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LITHOSPHERE 2004

THIRD SYMPOSIUM ON
THE STRUCTURE, COMPOSITION AND EVOLUTION
OF THE LITHOSPHERE IN FINLAND

PROGRAMME AND EXTENDED ABSTRACTS

edited by
Carl Ehlers, Olav Eklund, Annakaisa Korja, Annika Kruuna, Raimo Lahtinen, Lauri J. Pesonen

Åbo Akademi University,
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LITHOSPHERE 2004

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Keywords (GeoRef Thesaurus, AGI): lithosphere, crust, upper mantle, Fennoscandia, Finland, Precambrian, Baltic Shield, symposia

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PREFACE

The tremendous amount of new seismic reflection profiles from the deep seismic ***FIRE*** project will give us new and unique possibilities to study the structure and evolution of the lithosphere in Finland and will certainly keep both geophysicists and geologists busy for some time forward. One aim of the first Lithosphere meeting in 2000 was to increase the contact surface between geologists and geophysicists and to act as a forum for scientific correlation of surface geology and sub-surface geophysics. This is still the major aim of the present third biennial ***LITHOSPHERE 2004*** symposium after the successful two earlier meetings. The earlier trend of studies concerning the more recent evolution and activities within our Fennoscandian Shield seems to continue, and it is our wish in the ***ILP*** Finnish National Committee that the symposium this time again will give a picture of the state of the art within our fields of research. We hope that ***LITHOSPHERE 2004*** will promote more cooperation between geophysical and geological studies of the lithosphere and that it will stimulate further modelling and new research projects.

The two-day symposium will this time take place in the Arken campus, Åbo Akademi University, Tehtaankatu 2, Turku, November 10-11, 2004 with participation from the Universities in Helsinki, Turku and Oulu, the Geological Survey of Finland and the Finnish Geodetic Institute. The Symposium will be hosted by the ILP and supported by POSIVA Oy and Stiftelsen för Åbo Akademi. Posters prepared by graduate- or postgraduate students will be evaluated and the best will be awarded.

This symposium has the following themes:

Theme 1. The Structure and Composition of the Crust
Theme 2. Three-dimensional Models of Fennoscandia
Theme 3. Crustal Movements
Theme 4. Deep Biosphere and IODP
Theme 5. Initiating New ILP Proposals
Theme 6. Open Forum
Theme 7. General Discussion and Poster Awards

This special volume ***"Lithosphere 2004"*** contains the programme and extended abstracts of the symposium in alphabetical order.

Turku, October 28th, 2004

Carl Ehlers, Olav Eklund, Annakaisa Korja, Raimo Lahtinen, Lauri J. Pesonen
Organizing Committee

PROGRAMME

Wednesday, November 10, 2004, 10.00-19.00

09.30 – 10.00 Registration at Åbo Akademi University, ARKEN campus Tehtaankatu 2, Turku

10.00-11.00 Opening Session
Chair Carl Ehlers

10.00.10.05 Opening of the symposium by the Organising Committee (*Carl Ehlers*)
10.05-10.20 One more time: SVEKALAPKO; some seismological comments. (*S-E. Hjelt*)
10.20-11.00 FIRE transects: New images of the Precambrian crust. *(Heikkinen, P., Kukkonen, I.T., Ekdahl, E., Hjelt, S.-E., Korja, A., Lahtinen, R., Yliniemi, J., Berzin, R. and FIRE Working Group)*

14.50-15.15 Coffee Break

14.50–12.25 Three-dimensional Models of Fennoscandia
Chair Ilmo Kukkonen

11.10-11.35 3-D geophysical crustal model of Finland. (*Pirttijärvi, M., Kozlovskaya, E., Elo, S., Hjelt, S-E., Yliniemi, J.*)
11.35-12.00 Seismic tomographic images of the crust in southern and central Finland. (*Hyvönen, T., Tiira, T., Rautioaho, E., Korja, A., Heikkinen, P.*)
12.00-12.25 Structure of the upper mantel beneath central Fennoscandian Shield from seismic anisotropy studies. *(Kozlovskaya, E., Vescey, L., Plomerová, J., Babuska, V., Hjelt, S.-E. and Ratia, T.)*

12.25-13.30 Lunch break

13.30-15.10 The Structure and Composition of the Crust, Part I
Chair Pekka Nurmi

13.30-13.55 The processes forming the Palaeoproterozoic Svecofennian. (*Korja, A., Lahtinen, R., Nironen, M.*)
13.55-14.20 Thermal modelling of crustal stacking and exhumation during the Palaeoproterozoic orogenic growth of the central Fennoscandian Shield. (*Kukkonen, I.T., Korja, A., Lahtinen, R., Heikkinen, P., & FIRE Working Group)*
14.20-14.45 Isotopic constraints for the origin of the mafic lower crust at the Karelian Craton margin. (*Peltonen, P., Mänttäri, I., Huhma, H.*)
14.45-15.10 Orogenic architecture and mineralization - general concepts, comparisons and relevance to Fennoscandia. *(Sorjonen-Ward, P.)*

15.10-15.30 Coffee Break

15.30-17.00 Crustal Movements
Chair Krister Sundblad

15.30-15.55 The Lake Ladoga basin: preliminary insights into geochronology, igneous evolution, and tectonic significance.
(*Rämö,O.T., Mänttäri, I., Kohonen, J., Upton, B.G.J., Vaasjoki, M., Luttinen, A.V., Lindqvist, V., Lobaev, V., Cuney, M., Sviridenko, L.P.*)

15.55-16.20 Middle Proterozoic-Palaeozoic tectono-thermal reactivation of the crust in southern Finland and Northwestern Russia – paleomagnetic evidences.
(*Mertanen, S.*)

16.20-16.40 The Fennoscandian Land Uplift Gravity lines: Status 2004.
(*Mäkinen, J., Engfeldet, A., Harsson, B.G., Ruotsalainen, H., Strykowski, G., Oja, T., Wolf, D.)*

16.40-17.00 BIFROST: Continuous GPS measurements of the three dimensional deformation in Fennoscandia.
(*Koivula, H., Poutanen, M., Johansson, J.M., Scherneck, H-G., Milne, G.A., Mitrovica, J.X., Davis, J.L., Vermeer, M.*)

17.00-19.00 Poster Session
Chair Pekka Heikkinen

P1 Carbonate vein-dykes in Naantali: a new occurrence of carbonatite in SW Finland? (*Woodard, J., Hölttä, P.*)

P2 Petrogenesis of charnockites in the Turku granulite area. (*Helenius, E-M., Eklund, O., Väisänen, M., Hölttä, P.*)

P3 Single zircon U-Pb age results from the late Svecofennian microcline granites, southwetern Finland. (*Kurhila, M., Vaasjoki, M., Mänttäri, I., Rämö, T., Nironen, M.*)

P4 Late-Svecofennian deformation – evidence from D4 structures from the Kemiö island, SW Finland. (*Engström, J., Levin, T.*)

P5 The Anjalankoski earthquake swarm in May 2003.
(*Uski, M., Korja, A., Elo, S.*)

P6 Integration of geophysical data with sediment composition, examples from Prydz Bay (ODP Site 1166) Antarctica. (*Tiensuu, K., Strand, K.*)

P7 Clay mineral Occurrence in Prydz Bay Rize (ODP Site 1165) Antarctica: Implications for the Middle Miocene and the Middle Pliocene ice sheet evolution. (*Siira, T., Junttila, J., Strand, K.*)

P8 World Digital Magnetic Anomaly Map - presenting lithospheric contribution to the Earth's total magnetic field. *(Korhonen, J.V., Reeves, C., Ghidella. M., Maus. S., McLean. S., Ravat. D.)*

P9 Oxygen isotope and trace element zoning in garnets from granulite facies psammopelitic migmatite, SW Finland. *(Nyström, A.I., Whitehouse, M.J. Kriegsman, L.M.)*

19- ***Conference Dinner***
Organized by Pulterit ry at the “Gadolinia” building, Porthaniankatu 3.

Thursday, November 11, 2004, 9.00 – 1700

09.00– 10.20 Deep Biosphere and IODP
Chair Olav Eklund

09.00 - 9.40 ***Key Note Lecture:***
Exploration of the deep subterranean biosphere. *(Pedersen, K.)*

09.40-10.20 IODP investigates the solid earth's cycles and global environmental change. (*Strand, K.*)

10.20 –10.40 Coffee Break

10.40-12.00 The Structure and Composition of the Crust, Part II
Chair Raimo Lahtinen

10.40-11.05 Timing of crustal growth and collissional tectonics in Åland SW Finland. (*Ehlers, C., Skiöld, T., Vaasjoki, M.*)

11.05-11.30 Tonalites and similar rocks formed in pre- to syn-collissional stages of the Svecofennian orogeny. (*Eklund, O., Jurvanen, T., Väisänen, M.*)

11.30-11.55 Composition of the crust in the central Fennoscandian Shield: Lithological modelling of seismic velocity data.
(*Kuusisto, M., Kukkonen, I., Heikkinen, P., Pesonen, L.J.*)

12.00-13.00 Lunch

13.00 -14.00 Initiating New ILP Proposals – A Brainstorm Session
Chair Ilmo Kukkonen

14.00-14.50 The Structure and Composition of the Crust, Part III
Chair Annakaisa Korja

14.00-14.25 New zircon and monazite ages and Nd isotope data on the Arhean complex in Koillismaa, eastern Finland. (*Lauri, L.S., Huhma, H., Mänttäri, I., Räsänen, J.*)

14.25-14.50 Sm-Nd isotopes in Palaeoproterozoic mafic rocks in Finland –rifting of Archean lithosphere and mantle sources. (*Huhma, H., Hanski, E., Vuollo, J.)*

14.50-15.15 Coffee Break

15.15-16.00 Open Forum, Short Presentations (5-10 min)
Chair Sven-Erik Hjelt

ICDP *(Kukkonen, I.T. et al.)*

IGCP –projects

A research project: An integrated 3-D geophysical model of the lithosphere in the central Fennoscandian Shield (*Kaikkonen et al.,)*

Stabilization process of the Precambrian continental crust (*Kosunen, P. et al.)*

Proterozoic mantle enrichment from Ladoga to Juankoski (*Shebanov, A. et al.)*

16.00-16.30 Closing Session
Chair Olav Eklund
Poster Awards
General Discussion
Concluding Remarks

EXTENDED ABSTRACTS

EXTENDED ABSTRACTS

Timing of Svecofennian crustal grovth and collissional tectonics in Åland, SW Finland

Carl Ehlers[1], Torbjörn Skiöld[2] and Matti Vaasjoki[3]

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In an effort to quantify the timing of the tectonomagmatic evolution of the Southern Svecofennian Arc Complex (SSAC), advanced radiometric techniques have been applied on structurally distinct rock complexes. We report the results of high-resolution ion micro probe spot analyses (SIMS) on zircon determined at the Nordic NORDSIM facility and measurements using isotope dilution mass spectrometry (ID-TIMS) on zircon, monazite and titanite applying both large-sample multi-grain as well as single-grain techniques (Ehlers, Skiöld, Vaasjoki 2004)

A collage of imbricated rock slabs, the result of tectonic shortening, representing the culmination of large-scale penetrative deformations in the Svecofennian has been investigated in the Torsholma area, Åland Islands. We consider this to be a key area of structural significance for resolving some important phases in Svecofennian tectonics.

Another structurally significant feature investigated is the South Finland Shear Zone (SFSZ), that transects the southwestern archipelago and follows the south coast of Finland. This shear zone forms the southern limit of the c. 1830 Ma Late Svecofennian Granite and Migmatite zone and also features deformations of a later stage when the southern region of the exposed Svecofennian was consolidated.

The accumulated age results and tectonic analyses may be summarized as follows. The Enklinge volcanic sequence (1885 ∂ 6 Ma) is within error coeval with the intrusion of the abundant early-kinematic gneissose granodiorites whose average age of 1884 ∂ 5 Ma marks the formation of new crust in this region. Some of these gneisses contain a significant amount of 2000 - 2080 Ma zircon. Altough many Svecofennian granitoids are known to contain heterogeneous (inherited) zircon populations, the Kökar gneiss is, to the best of our knowledge, the first case were inheritage from c. 2030 Ma sources has been unequivocally demonstrated in a syntectonic Svecofennian intrusive rock.

At Torsholma, the granodioritic orthogneisses (1879∂ 6 Ma) similarly intruded the supracrustal series, and subsequently became metamorphosed into granulites. The mesoscopic recumbent folds and subhorizontal schistosities of these gneisses were transected by a set of steep amphibolitic dykes indicating an episode of extension. After this phase the gneisses were sandwiched between the supracrustal rocks, and subsequently intruded by granodioritic sheet-like intrusions (1861∂19 Ma) with associated dykes (1865∂7 Ma). This tectonic scenario took place during approximately 15 Ma (c. 1875 – 1860 Ma) and records the culmination of the Svecofennian collissional deformation.

Monazites and zircon overgrowth textures give concordant U-Pb ages of c. 1830 Ma in accordance with previous ages on surrounding late Svecofennian granites and migmatites of the LSGM zone in S Finland.

A steep undeformed pegmatite dyke with a monazite age of 1795±4 Ma transects the whole set of imbricated rock slabs in Torsholma (Fig. 2). The overlapping U-Pb ages of this and other pegmatites of the post-kinematic type intrusions, as well as the timing of the large-scale shear zones, indicate consolidation of the Svecofennian crust. This also indicates that deformation at this time was partitioned along steep discret crustal shears in contrast to the earlier subhorizontal crustal shortening and thickening seen in the Torsholma area. The transition between the earlier tectonic history of subhorizontal collissional crustal deformation, and the later phases of strain partitioning into steeply dipping shears must have happened after the intrusion of the subhorizontal granite sheets (c. 1830 Ma) belonging to the LSGM zone of southern Finland.

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Ehlers, C., Skiöld, T-B., Vaasjoki, M. 2004. Timing of Svecofennian crustal grovth and collissional tectonics in Åland, SW Finland. Bull.Geol.Soc.Finland, vol. 76 (in print)

Tonalites and similar rocks formed in pre-to syn-collisional stages of the Svecofennian orogeny

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Tonalites are the pricipal continent forming rocks. Our preliminary results show that the tonalites in the accretionary arc complex of southern Finland may have been formed by several petrogenetic processes over a period of 80 Ma years (1900 Ma to 1820 Ma).

1) 1900 Ma tonalites and gabbros associated with volcanic rocks of the accretionary arc-complex of southern Finland. These tonalities can be classified as M-type granitoids.
2) The 1880 Ma Enklinge intrusion is a bimodal tonalite-gabbro intrusion with geochemical character of a typical syn-collisional rock series.
3) In Turku area (and areas in eastern Finland) tonalites evidence an adakitic geochemical signature, that may correspond to melting of a metabasalt at high (15 kbar) pressure. This metabasalt may represent a subducting plate.
4) In some areas of southern Finland melting of amphibolites to tonalities takes place due to high-T low-P regional metamorphism.

Keywords: Proterozoic tonalites, adakites

Compared to the late-, post- and anorogenic Svecofennian magmatism, the synorogenic magmatism has been little studied in Finland and Sweden. Even though tonalites are the principal crust forming rocks, very little effort have been made to study these rocks in detail in the proterozoic part of the Fennoscandian shield. From the accretionary arc complex of southern Finland there are studies by *Arth et al. (1978)* and *Nurmi and Haapala (1986)* that deal with petrogenetic questions on these rocks. Tonalites have been described from a metamorphic point of view in southern Finalnd, where tonalite bodies appear in both amphibolite and granulite facies *(Mäkitie 1993; Väisänen and Hölttä, 1999; Väisänen et al., 2002; Helenius, 2003; Helenius et al. this volume)*.

During the last few years the Department of Geology at Turku University has collected data about the syn-orogenic magmatism and tonalite genesis from the accretionary arc complex of southern Finland. These studies have shown that the rocks that have generally been mapped as quartz-diorites, tonalites, trondhjemites and granodiorites in southern Finland were formed over a time span on 80 Ma through several petrogenetic processes. In this text the term tonalite is used to describe this group of rocks.

The study by *Väisänen (2002)* indicate that the tonalite associated with the earliest extrusive volcanic arc rocks in the Orijärvi area are approx. 1900 Ma. This age is contemporaneous with the age of volcanism in the are, why this tonalite may be considered as a pre-collisional tonalite or as defined by *Pitcher (1993)* a M-type granite.

The studies by *Impola (2000)* and *Eklund (2002)* indicate that the 1880 Ma *(Suominen, 1991)* quartz-diorite and tonalite in Kumlinge (Åland archipelago) seem to be magmatic differentiates from a more gabbroic magma, similar to that described by *Arnt et al. (1978)* from the Kalanti area in SW Finland. The geochemical data in the thesis by *Helenius (2003)* indicate that tonalities in the Turku area have high Na_2O, Mg#, Sr/Y and La/Yb, geochemical signatures for adakitic rocks. This means that these tonalites may have been formed by partial melting of a meta-basalt at high pressure (around 15 kbar). This metabasalt may be a subducting slab. In figure 1 geochemistry from Enklinge tonalities and Turku tonalites are compared. The figure show that the two areas show different geochemistry with some overlappings.

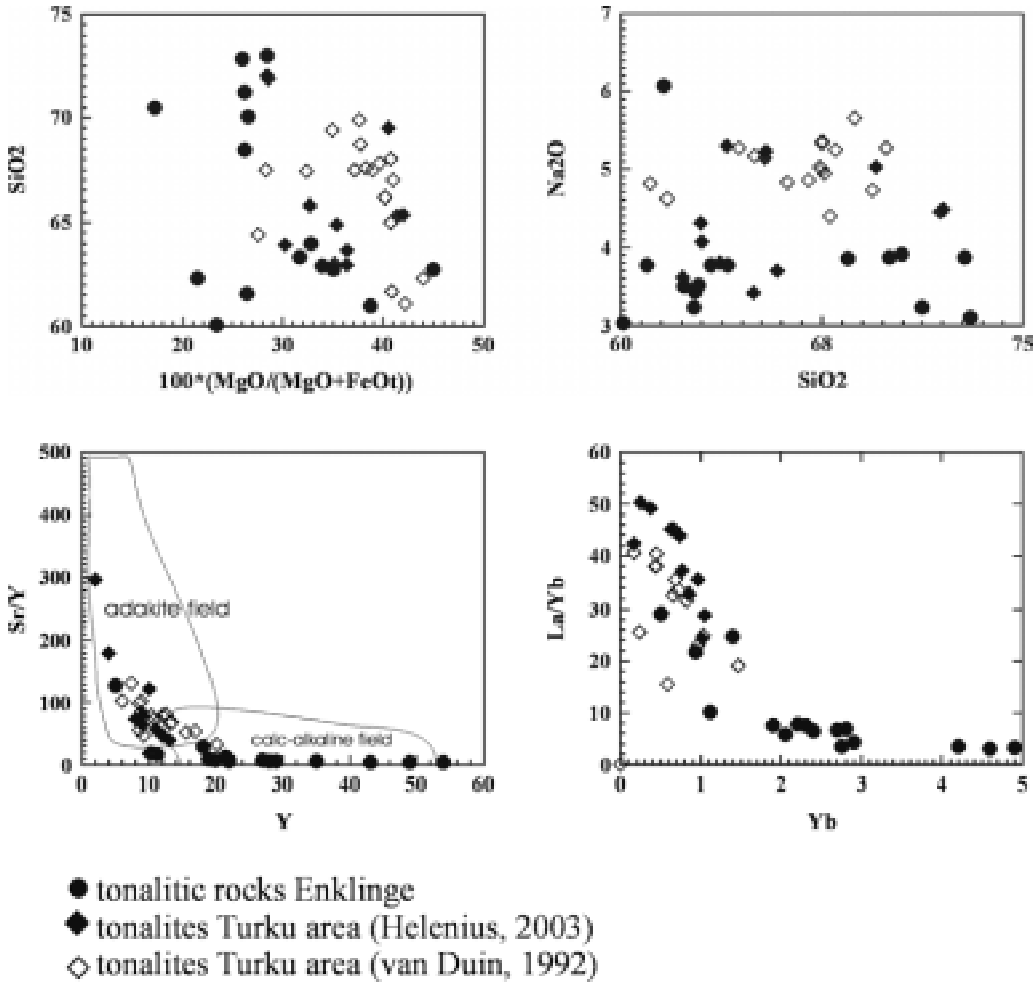

Figure 1a. SiO2 versus Mg-number indicating that the tonalites from Turku are have higher Mg-number compared to tonalites from Enklinge area.

Figure 1b. Diagram showing higher Na2O contents for the tonalites in Turku area compared to Enklinge area.

Figure 1c. Sr/Y versus Y diagram that discriminate adakitic tonalites from calk-alkaline tonalites. Tonalites from Turku area plot in the adakite field and tonalites from Enklinge in the calk-alkaline field.

Figure 1d. La/Yb versus Yb diagram indicating that tonalites from Turku area have high La/Yb ratio and low Yb contents, typical for adakitic rocks.

Lepistö (2003) and *Naumanen (2003)* investigated a tonalite at Höggrund (western Åland archipelago) that was associated with amphibolitic enclaves. The results of their study was that the tonalite was formed during the regional high T, low P metamorphism in southern Finland at 1840-1820 Ma by melting of amphibolites leaving an amphobile rich melanosome. Amphibolite melting experiments (samples from SW Finland) was undertaken by Lindberg *(2002)*. He concluded that melting of amphibolites took place during the regional metamorphose in southern Finland due to excess of fluids in the system.

Observations of partial melting of amphibolite to tonalite as a consequence of the regional high-T Low-P metamorphism (at ca 1830 Ma) have been made by Kriegsman (1999) and Lindberg (2002). The regional metamorphism did also affect the tonalites, particularly in Turku area, the tonalites were objects to charnockitization at roughly 1820 Ma (Väisänen et al., 2002; Helenius, 2003).

These data show that we about tonalites can use a modified phrase initiated by H.H. Read, “there are tonalites and tonalites” (originally “there are granites and granites”).

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Late-Svecofennian deformation – evidence from D_4 structures on the Kemiö Island, SW Finland

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The D_1 structures in the study area on the Kemiö Island are mostly small-scale and transposed and overprinted by D_2 structures. The D_3 deformation phase can be seen as open folds at a regional scale and thus visible only indirectly in the outcrop pattern. An S_4 crenulation cleavage in the quartz schist, small F_4 folds in the marble and an S_4 foliation in the sillimanite schist represent the late-Svecofennian deformation in the study area. These structures have a constant angle (ca. 25°) against the predominating S_2 schistosity. As a result of the obtained structural and petrological data a tentative 3D-model of the geological structure is presented.

Keywords: Late-Svecofennian, structural geology, deformation, crenulation cleavage, sillimanite gneiss, quartz schist, Kemiö, SW Finland

The Kemiö Island in SW Finland and the studied Bjensböle area within the migmatite zone of southern Finland are rich in supracrustal rocks that expose evidence of the main deformation phases (D_1-D_4). The supracrustal rocks are covered by sheets of 1,84-1,81 Ga old microcline granite in the northern and southern parts of the island *(Seitsaari 1955; Dietvorst 1982).* The rocks in the study area consist of low P/high T amphibolite facies sillimanite gneiss and low P/low T amphibolite facies quartz schist, separated by a metavolcanic amphibolite horizon in the central part of the area. Thin horizons of marble and black schist have been used as marker horizons.

Three major phases of deformation has been recognized in SW Finland *(Schreurs & Westra 1986; Ehlers et al. 1993; Selonen & Ehlers 1998; Väisänen 2002).* The oldest structure in the study area, S_0, is defined as the original bedding of the supracrustal rocks. The bedding is only seen at a few localities. The D_1 structures are mostly small-scale and transposed and overprinted by D_2 structures. The F_2 folds are tight to isoclinal with sub-horisontal axial planes. The D_2 phase also exhibits a penetrative axial plane foliation (S_2).

SW Finland can be divided into two major geological and structural domains, the Kemiö-Orijärvi domain and the Turku domain. The Paimio shear zone separates these two domains from each other. The D_3 deformation phase is well exposed in the Turku domain with visible S_3 schistosities and small-scale F_3 folding *(Väisänen & Hölttä 1999).* In the Kemiö-Orijärvi domain the D_3 deformation phase only exhibit open folds at a regional scale; the D_3 deformation phase seldom exhibits an own schistosity or small-scale folds and is thus visible only indirectly in the outcrop pattern. The mesoscopic structures studied in outcrops are thus predominantly D_2 and D_4.

A deformed and folded marble horizon in Bjensböle was studied. The marble layer dips gently towards N and is situated in the southern leg of an open regional F_3 fold. The rocks on both sides of the marble are tilted and almost horizontal, with shallow dips towards the north. The S_1+S_2 schistosities in the marble are folded into open to tight asymmetric folds (wavelength 5-10 cm) plunging ca. $32^o/20^o$. As these folds post-date and refold the S_2 axial plane foliation and also refold the southern flank of an F_3 fold, they have been identified as F_4 folds representing the D_4 deformational phase. The D_4 phase is also visible in the quartz schist area. At a few localities the F_2 folds have been refolded by D_4, creating new small-scale F_4 folds and a new S_4 axial plane schistosity (fig. 1).

Figure 1. Horizontal photo showing an intrafolial F_2 fold and the D_4 crenulation cleavage in the quartz schist.

Two generations of foliations are found in the sillimanite gneiss, one defined by biotite (S_1+S_2) and the other by sillimanite (S_4). The sillimanite has grown along the S_2 foliation plane of the micas, but has an overprinting direction of their own. This overprinting schistosity can also be seen at a smaller scale in thin sections. The S_4 crenulation cleavage in the quartz schist, the small F_4 folds in the marble and the S_4 foliation in the sillimanite schist all have a general NE direction and a steep dip. The angle against the predominant E-W striking S_2 schistosity (ca. 25˚) is also nearly constant.

As a result of the obtained structural and petrological data a tentative 3D-model of the geological structure showing the D_2 and D_4 deformation in the study area is presented (fig. 2). The D_2 deformation exhibits an overturned tight F_2 fold and the F_4 folds are exposed both at a macroscopic and mesoscopic scale. The macroscopic folds (F_{4a}) are open and extend over the entire study area. The smaller mesoscopic folds (F_{4b}) include the crenulation cleavage and the marble folds and are restricted to the southern part of the study area. This implies that the D_4 phase has deformed all the rocks in the study area, resulting in an approximately NE-SW striking axial plane direction, indicating a NW-SE directed compression for the D_4 deformation. The structural evidence point to a dextral sense of regional shearing for the D_4 deformation phase, possibly related to regional ductile shearing signifying the vaning stages of the late-Svecofennian transpressive deformation.

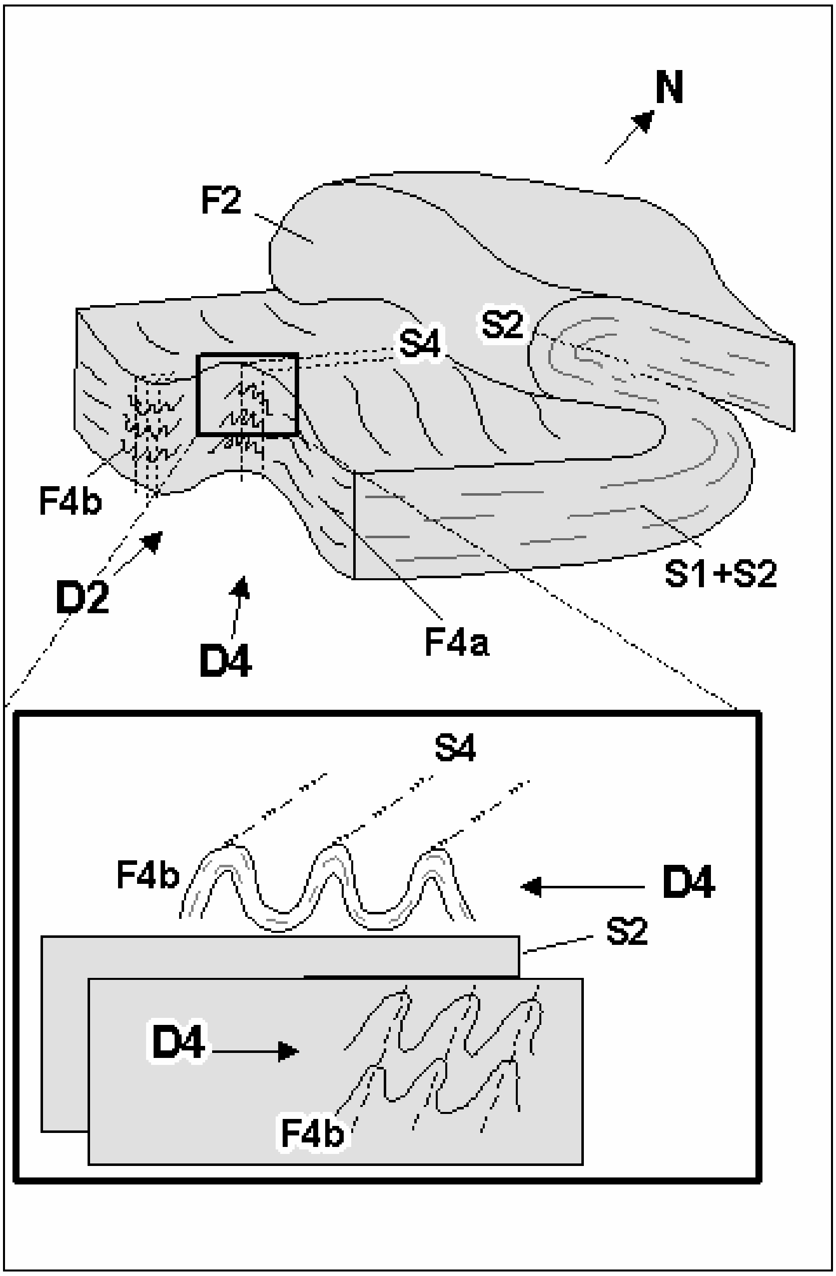

Figure 2. Simplified schematic 3D-model of the study area showing an overturned F_2 fold refolded by the D_4 deformation phase. The D_4 deformation phase exhibits folds in mesoscopic (F_{4a}) and microscopic (F_{4b}) scale. The open F_3 structures are omitted from the model.

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FIRE transects: New images of the Precambrian crust

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The FIRE (Finnish Reflection Experiment) consortium carried out Vibroseis measurements in the Fennoscandian Shield in 2001-2003 on four lines with a total length of 2135 km and transecting the major geological units on the Shield. The seismic reflection data provides new and in many regions unexpected images of the crustal structure of the Fennoscandian Shield. In the future, they will be an indispensable component of the studies of the tectonic evolution of the Shield.

Keywords: reflection seismics, crustal structure, Fennoscandia, Precambrian

1.Introduction

Seismic reflection profiling has turned out to be the most efficient tool in studying the deep structure of the earth's crust. The large scale reflection surveys conducted during the last 10-15 years in Canada in the Lithoprobe project (*Clowes et al., 1996*), in Australia (*Korsch, 1998*) and also in the Fennoscandian Shield (*BABEL Working Group, 1993a,b, Korja & Heikkinen, 2004*) have provided new insights on the structure and evolution of the Precambrian crust.

The FIRE (Finnish Reflection Experiment) consortium – consisting of the Geological Survey of Finland, and Universities of Helsinki and Oulu, with Russian company Spetsgeofizika S.G.E. as a contractor, carried out a total of 2165 km of Vibroseis measurements in the central part of the Fennoscandian Shield in 2001-2003. The four FIRE transects (Figure 1) run over the most important geological units of the Fennoscandian Shield. Where possible, the lines follow existing deep refractions profiles shot in 1981-1994 (*Luosto,* 1997 *and refrerences therein; FENNIA Working Group, 1998*).

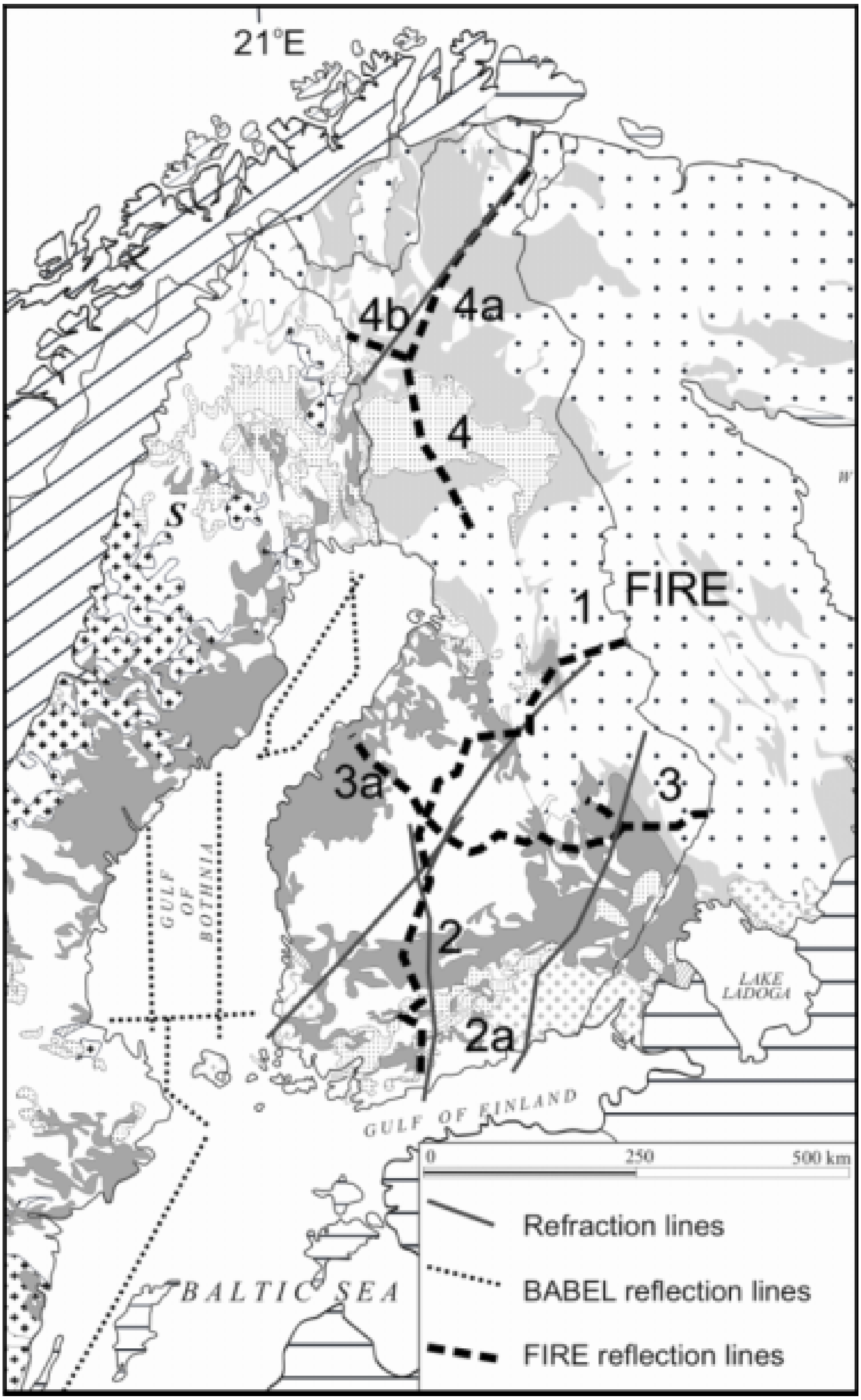

Figure 1. Seismic deep profiling lines in Finland on lithological map modified from Koistinen et al. (2001).

2. Data acquisition and processing

Line 1 runs from the Karelian craton in NE, crossing the Archaean-Proterozoic boundary and ends in the Central Finland Granitoid Complex (CFGC). Line 1 follows closely the deep refraction profile SVEKA. Line 2 is a direct continuation of line to the south along the refraction profile FENNIA. Line 2 begins in the CFGC, crosses the Tampere Schist Belt and terminates in the Southern Finland Arc Complex. Line 3 runs E-W from the Archaean in the east to the Bothnian Schist Belt in the west. The first part of the 570 km long FIRE 4 runs in S-N direction from the Archaean in the southern end of the transect, crossing the Proterozoic metasediments and the Lapland Granitoid Complex and ending at the Kittilä Greenstone Belt. The second leg of the line runs from the Kittilä Greenstone Belt in SW-NE direction crossing the Lapland Granulite Belt and ending on the Archeaean Inari Terrain. The second part (FIRE4a) coincides with the deep refraction profile POLAR, shot in 1985. A 60-km long side track (FIRE4b) was shot from Sirkka to Muonio at the Swedish border to provide 3-D information of the crustal structures under the Kittilä Greenstone Belt. In those parts of the lines where previous deep refraction data were no available - line 3 and southern part of line 4 - wide angle measurements were carried out utilizing the Vibroseis signal.

The frequency range of the Vibroseis sources used was 12-80 Hz corresponding to wavelengths of about 500-75 metres. The spacing of the source points was 100 m and receiver spacing along the 18 km long spread was 50 m. As a whole the FIRE data is of good quality. The most important quality controlling factors were optimal selection of sweep parameters, careful geophone installation, and continuous supervising and control by the client. Difficult conditions and local deterioration of data quality were encountered in sections with soft and thick Quaternary soil types (peat, clay, silt and sand). Acquisition conditions were most favourable in areas with a thin overburden of till, and where the acquisition was carried out during season when frost had penetrated deep in the sediments (early spring).

The basic data processing was done by Spetsgeofyzika producing two data sets: normal CMP-stack with record length of 30 s which corresponds about 100 km depth and 16 s long DMO-stacks in which steeply dipping structures are imaged better. The sections were afterwards migrated at the Institute of Seismology.

3. Results

The upper crust on all FIRE profiles is generally very reflective and numerous bright reflections can be traced from the surface to the middle crust. These are mostly due to collision and thrusting events during the Svecofennian orogeny (about 1.9-1.8 Ga). Correlation of surface geology and geophysics has provided direct links to the geological characteristics of upper crustal reflections. In most cases where direct comparison of reflections and outcrops is possible, the upper crustal reflections seem to represent mafic or intermediate rocks in a more felsic background.

3.1. FIRE 1, 2 and 3

The first transect, FIRE-1, images the break-up of the Archaean craton and opening of the Svecofennian Sea, the subsequent closure of the oceanic basin, and collisions of island and continental arc complexes against the craton margin. The Archaean upper and middle crust, in the margin, is very reflective due to Proterozoic mafic magmatism related to multiple rifting and break-up. Later collisional thrust surfaces at the margin are seen as strong continuous reflections, which can be followed from the surface to the depth of c. 15 km. In the southwestern part of the transect, N dipping strongly reflecting zones, interpreted as continental stacking structures, were discovered. Sharp, sub-horizontal and very continuous reflections on FIRE 1 and FIRE 2 at about 2 s TWT beneath the Central Finland Granitoid Complex possibly represent buried volcanic units or younger diabase sills.

A completely unexpected result of FIRE is the very continuous eastward dipping reflections recorded on FIRE 3 under the central Finland Granitoid Complex and the surrounding schists belts. The reflections are extending up to the margin of the Archaean craton in the east. These reflections which dominate the crustal scale image on FIRE 3, continue from the surface at least to the base of the upper crust, and possible even to the base of middle crust at about 40 km depth. Their origin is not yet understood, but these reflections could be attributed to stacking of crust during the Svecofennian orogeny at about 1.90 Ga ago, or to post-collisional extension. The crossing of FIRE 1 and 3 in western Finland will allow viewing of the reflecting structures in 3-D.

The FIRE results indicate that the reflection Moho is very deep and it is in a good agreement with the previous wide-angle studies. The reflection Moho is between the depths of 52km and 55 km, and it is diffuse in nature. The lower crustal reflectivity ends over a depth interval of about 3-5 km. The reflection Moho is gently undulating, and sharp variations in crustal thickness are usually not seen. However, near the eastern end of the FIRE 1, a sudden step-wise thinning of crust of about 15 km is seen under the Archaean craton. This result is supported by previous wide-angle studies. This jump is attributed to either collisional stacking of the crust or to tectonic underplating during the Svecofennian orogeny.

The lower crust mostly lacks long continuous reflections, and instead it comprises short (about 1 km long) and thin (<0.5 km thick) reflections. Strong, dipping reflections dominating the reflection image in the upper and middle crust have only weak or counterparts in the lower crust. Sharp reflection Moho was recorded only at two localities. The first one is at the western end of FIRE 3, close to a major crustal boundary between the Central Finland Granitoid Complex and the Bothnian Schist Belt, where the diffuse lower crust in the east changes into a laminated brightly reflective lower crust in the west across a sub-vertical crustal scale discontinuity. The second locality with reflective lower crust was recorded beneath the Lapland Granulite Belt on FIRE 4.

3.2. Fire 4

In general, the tectonic units along the FIRE 4 transect have distinct reflectivity patterns and especially in the lower crust quite different form those in the lines 1,2 and 3. The craton margin at the southern end is dominated by strong reflectivity in the upper 20 km-30 km, probably due to mafic magmatism related to the break-up of the craton and opening of the Proterozoic Peräpohja Basin. Except for the first few kilometres, the metasedimentary units and the southern part of the Lapland Granitoid Complex are characterized by weaker reflectivity. The nature of reflectivity changes clearly in the middle of the Lapland Granitoid Complex. In the northern part of the Granitoid Complex the upper and middle crust to the depth of 20 km are dominated by strong reflectivity.

The Kittilä Greenstone Belt is shown as a weakly reflective unit that is about 10 km thick. The reflective structures at the belt boundaries are well correlated with the previously known tectonic structures such as the Sirkka Line in Central Lapland, a major discontinuity in lithology and metamorphic grade. This S-dipping thrust surface can be followed to the depths of 10-20 km. The southwestern part of the Lapland Granulite Belt is visible as a zone of strong reflectivity whereas the northern part of the belt is weakly reflective. The belt as a whole is a wedge-shaped unit with maximum thickness of 15 km.

The Moho boundary is clearly visible along the whole FIRE 4 at the depth range of 40-50 km and differing markedly from the other FIRE lines. Beneath the Archaean craton, the thickness of the crust is 42 km, thickening to 47-48 km beneath the Central Lapland Granitoid Complex, thinning back to 40-42 km beneath the Lapland Granulite Belt and thickening again to 50 km at the northern end of the line within the Archaean Inari Terrain. In the velocity model of the POLAR profile, the thinner crust beneath the Lapland Granulite Belt is characterized by lower velocities in the lower crust. In the reflection data this part is shown as a strongly reflective laminated zone extending beneath the Inari Terrain.

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Petrogenesis of charnockites in the Turku granulite area

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Charnockites are either plutonic or metamorphic orthopyroxene-bearing quartzofeldspathic rocks. This study shows that the charnockitic rocks within the Turku granulite area were formed under granulite facies conditions from tonalitic granitoids through charnockitization reactions. Biotite and amphibole reacted with quartz to form pyroxene, plagioclase ± potassium feldspar.

Keywords: charnockitization, granitoid, metamorphism

1. Introduction

Charnockites are orthopyroxene-bearing quartzofeldspathic rocks, whose origin can be either plutonic or metamorphic. The term "charnockite" is by Sir Thomas Holland, who created the term in the end of the 19th century originally to mean only plutonic orthopyroxene-bearing rocks in India. The name was given after Job Charnock, the founder of Calcutta, whose tombstone is made of this type of rock. Nowadays the nomenclature of orthopyroxene-bearing felsic rocks created by Streckeisen (1974) and Le Maitre (*1989*) is used.

Charnockites are located mainly in the areas of high grade metamorphism. They are known in almost every continent, particularly in the Precambrian areas (*Yardley 1989*). Zhao et al. (*1997*) lists several models proposed for their origin: "intrusion of high-temperature, water-deficient felsic magmas into dry granulitic crust, in-situ dry anatexis during granulite metamorphism, residua after removal of granitic melts, cumulates of igneous fractionation, partial melting of granulitic mineral assemblages enhanced by CO_2-fluxing, second-stage partial melting of melt-depleted granulitic source, crystal fractionation and crustal assimilation of mantle-derived gabbroic magmas, dehydration melting of dry Ti-enriched mafic underplates compositionally similar to high-Ti continental flood basalts". The most popular model proposed for their origin is the dehydration of granitic to granodioritic rocks in granulite facies metamorphism. This means that the water-bearing minerals (micas, amphiboles) react to form dry minerals (pyroxenes, garnets) in high grade conditions.

Orthopyroxene is very characteristic mineral in charnockitic rocks. Pertite, mesopertite or antipertite are also common. Brownish grey, grayish yellow or greenish yellow tone is also characteristic of fresh charnockitic rocks. Rust-coloured orthopyroxene patches can usually be seen on the weathered surface (*Parras 1958*).

The earlier investigations of charnockites in the Turku granulite area are mainly of Hietanen (1947) and van Duin (1992). Hietanen made a petrological interpretation of charnockites in the Turku area. Her interpretation was that charnockites are magmatic. Van Duin's (*1992*) interpretation of the pyroxene-bearing rocks in the Turku area was that the rocks of tonalite-trondhjemite-granodiorite (TTG) and charnockite series are plutonic and they were formed after differentiation of partial melts derived from the lower crust. The protolith for the TTG series rocks would have been amphibolite and for the charnockite series rocks garnet-bearing amphibolite. Väisänen et al. (*2002*) dated the zircons from the Turku charnockite with ion-probe. The cores gave ages of ca. 1870 Ma and the rims ca.

1820 Ma. The former corresponds to the age of the syn-orogenic granitoids in southern Finland and the latter age corresponds well to the age of metamorphism in the area.

The aim of this study is to investigate the petrogenesis of charnockitic rocks in the Turku area by mineralogical methods. The study was focused on the pyroxenes, whose magmatic vs. metamorphic origin was investigated. The research area is situated in the SW Finland, around the city of Turku. Samples were taken from the rocks in four different intrusions in Turku, Lieto, Masku and Naantali to study their whole rock chemistry, petrography and mineral chemistry. Whole rock analyses were made in Canada by ACTLAB (Activation Laboratories Ltd.) and mineral analyses were made in Geological Survey of Finland.

2. The geology of the Turku granulite area

The Turku granulite area is located within the Late Svecofennian Granite-Migmatite Zone (LSGM, *Ehlers et al. 1993*). It consists of supracrustal (pelitic) and infracrustal rocks. During the main period of the Svecofennian orogeny, 1.89-1.86 Ga ago, supracrustal rocks deformed and migmatised and the early Svecofennian granitoids (charnockite and TTG series) intruded into them mainly 1.88-1.86 Ga ago as plutons and sills. Charnockite series rocks are located mainly within the granulite facies area. Infracrustal rocks consist also of 1.83-1.81 Ga old microcline granites and post-tectonic c. 1.60 Ga old intrusions (diabase dikes, anorthosites, rapakivi granites).

High grade metamorphism in the southern Finland was caused by the thermal influence 1.83-1.81 Ga ago. Thermal influence caused the fluids to rise through the crust and the fluids influenced the rocks together with changes in temperature and pressure. Pressure and temperature estimates from garnet and cordierite bearing rocks indicate that the granulites reached temperatures in excess of 800 °C at approximately 6 kbar pressure while the adjacent amphibolite facies rocks crystallized at 650-700 °C and 4-5 kbar (*Väisänen & Hölttä 1999*). There is no tectonic boundary between the amphibolite and granulite facies domain (*Van Duin 1992*).

3. Results

Petrographic results show that the rocks are metamorphosed. The rocks are intermediate to felsic, homogeneous or deformed, granoblastic, and relatively even-grained. The grains are mainly anhedral and subhedral, except for small accessory minerals (apatite, zircon). The surface of the weathered rock is from light brown to dark grey and the colour of the fresh pyroxene-bearing rock is greenish. The rocks are metaluminous/peraluminous, calc-alkalic, island-arc–type granitoids. Six samples of ten are pyroxene-bearing. Three other samples of four contain hornblende and/or cummingtonite. The rocks plot in granodiorite, tonalite and quartz-diorite fields. Thus pyroxene-bearing rocks are enderbites, charnoenderbites and hypersthene-quartz-diorites.

TiO_2 content of biotite and hornblende in the pyroxene-bearing rocks is clearly higher than in the Hbl/Bt-bearing rocks (Fig. 1). Ti-content increases as temperature rises (*Henry & Guidotti 2002*) and its content helps to stabilise the biotite such that it persists throughout the metamorphic event (*Harlov & Melzer 2002*). This can indicate that the pyroxene-bearing rocks were formed as temperature rose in the Turku granulite area.

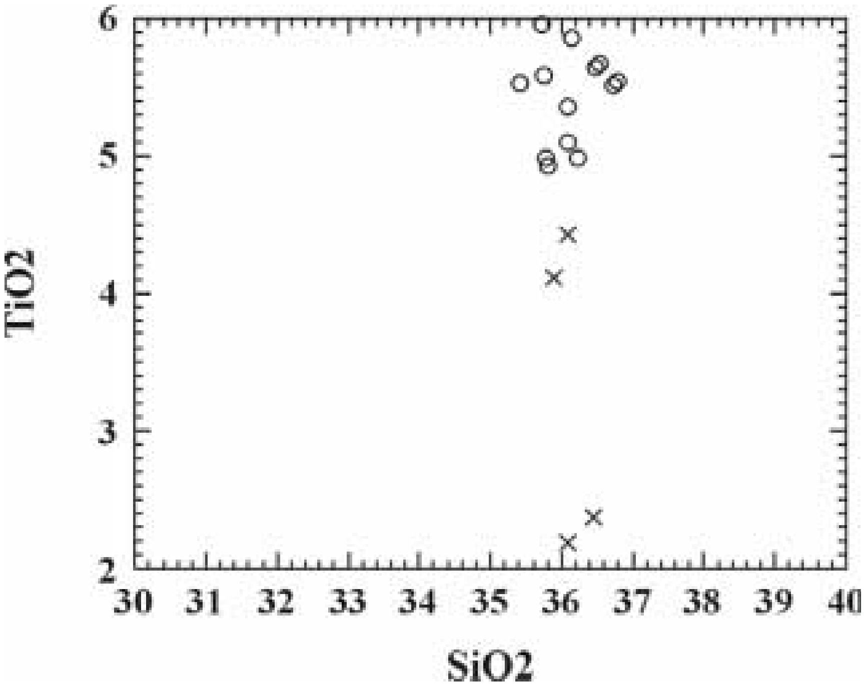

Figure 1. TiO_2 contents in biotites of the pyroxene-bearing rocks was clearly higher than in the Hbl/Bt-bearing rocks (pyroxe-absent). Circles are for biotite analyses from pyroxene-bearing rocks and crosses from pyroxene-absent rocks.

Feldspar content increased and biotite content decreased during metamorphism, as plagioclase was formed in the reaction where hornblende or biotite reacted with quartz to form pyroxenes and plagioclase. Pyroxene-bearing rocks contain more plagioclase than Hbl/Bt-rocks.

It is obvious that the structure of the rocks changes to more homogenous as the pyroxene content rises. This indicates that the charnockitization proceeded in some places further, and the rock was changed homogeneous and lost its cleavage. On the less charnockitizised areas the weak cleavage can be seen in the rocks and all the amphiboles have not reacted to form pyroxenes.

4. Conclusions

Charnockitic rocks in the Turku area were formed during granulite facies metamorphism from tonalitic granitoids after charnockitization reactions. In the dehydration reaction biotite and amphibole reacted with quartz to form ortho- and clinopyroxene, plagioclase ± K-feldspar and Ti-bearing biotite.

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One more time: SVEKALAPKO; some seismological comments

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The SVEKALAPKO-project, one of the five key project of the ESF-supported EUROPROBE programme, was operative during the years 1995 - 2001. Two of the major experiments performed during the active phase of SVEKALAPKO were the BEARstudy and the seismic tomography experiment. The analysis of the great wealth of data has been going on actively.
The paper summarizes once more some key results obtained so far from the tomography experiment. These comprise of tomographic slices of the ccntral Fennoscandian lithosphere, surface wave analysis, subcrustal anisotropy

***Keywords**:* Fennoscandian Shield, crust, upper mantle, seismic tomography.

1. Introduction

The key issue of the SVEKALAPKO-project (1995-2002) was to study the nature of the Fennoscandian lithosphere with 60km thick crust (*Hjelt et al., 1996*). The field campaign of two key experiments, the seismic tomography array and the deep electromagnetic BEAR array studies were successfully completed in 1998 and 1999 respectively and the analysis of the results has continued since then. The SVEKALAPKO project was summarized and the seismic tomography and BEAR subprojects were discussed in detail at the LITO2002 symposium (*Hjelt and Daly, 2002, Kozlovskaja et al, 2002b; T Korja et al, 2002*).

2. The seismic tomography experiment

The field work of the SVEKALAPKO deep seismic tomography experiment took place in 1998-1999. The first results not only corraborated, but also specified in more detail the known differences in the lithosphere beneath Fennoscandia and Central Europe (*Bock et al., 2001). The* increasing difference between the 410 and 660 km regions from TOR to SVEKALAPKO areas was a clear indication of a cold lithosphere of the Shield at depth. Further studies modelled the depth extension of a high-velocity keel below the central Fennoscandian Shield down to about 300 km. The upper mantle P-wave velocities differ a few % from the standard reference model IASP91. The analysis of subcrustal anisotropic properties seem to delineate the major geologo-tectonic units of the central shield. Crustal velocity models are continuously improved in order to increase the accuracy of upper mantle models below the SVEKALAPKO seismic tomography array. (e.g. *Alinaghi et al, 2003; Bock et al., 2001; Bruneton et al, 2002, 2004a, Kozlovskaya et al., 2002a,b , 2004 this issue; Plomerova et al, 2002; Sandoval et al., 2003)*

3. Future studies

The analysis of the seismic tomography data is still going on and much new details about the deep structure of the crust and upper mantle will certainly become available. Joint inversion of potential field and seismic tomographic data (*Kozlovskaya et al., 2004, Yliniemi et al., 2003)* have significantly improved the crustal velocity models. Important new constraints on the lithospheric models have been obtained through combined analysis of seismic surface waves and xenoliths (*Bruneton et al., 2004b*).

Still mote interesting results are to expected from seismic reflection data obtained during the Russian-Finnish FIRE experiment with its more than 2000 km of lines across key geological structures in the Finnish part of the Fennoscandian Shield. (e.g. *Heikkinen et al., 2004 and Kukkonen et al., 2004*)

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Sm-Nd isotopes in Paleoproterozoic mafic rocks in Finland – rifting of Archean lithosphere and mantle sources

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Isotopic studies on mafic dikes, intrusions and volcanic rocks have revealed that rifting of the Archean lithosphere took place at several stages, e.g., 2.45 Ga, 2.3 Ga, 2.2 Ga, 2.15 Ga, 2.1 Ga, 2.05 Ga, 2.0 Ga, 1.8 Ga. The initial ε_{Nd} values range from very positive to strongly negative and suggest that some rocks were derived from a depleted mantle source whereas others have a large contribution from old enriched lithosphere.

Keywords: Sm-Nd isotopes, mafic rocks, Fennoscandian Shield

1. Introduction

The range of ages of mafic rocks in the Fennoscandian Shield is well exemplified in the special volume containing vast U-Pb data and results from Northern Finland (*Vaasjoki, 2001*). The mafic rocks show a range of ages, mostly at c. 2.44 Ga, 2.21 Ga, 2.14 Ga, 2.05 Ga, and 2.0 Ga (*Hanski et al., 2001a*). Most of the U-Pb results date gabbroic rocks from intrusions and dikes, but can also be used to constrain the age of mafic volcanism, which has produced major formations especially in Lapland. Mafic rocks in a cratonic setting manifest rifting of the lithosphere and provide samples from the mantle. A large number of Sm-Nd mineral and whole-rock analyses have been made in order to constrain the age and origin of these rocks and the evolution of the lithosphere.

The emphasis has been on the Sm-Nd studies on most pristine mafic rocks available. Surprisingly well-preserved igneous rocks have been found after careful mapping and geological studies, even close to areas and zones of strong metamorphic reworking. The Sm-Nd mineral ages determined for such well-preserved rocks are generally consistent with the available U-Pb zircon ages. One exception is provided by the Rantavaara intrusion in Sodankylä, where the apparent old Sm-Nd mineral age is explained by slight crustal contamination in the intercumulus plagioclase, which is a minor phase in the ultramafic rock. Isotopic unequilibrium between clinopyroxene phenocrysts and whole rock may also cause problems in dating, as was observed e.g. in Suisarian picrites in Onega, Russia (*Puchtel et al., 1998; Huhma et al., unpublished*).

2. Results and discussion

The main results on the mafic rocks from the cratonic area are summarized in the ε_{Nd}-age diagram in Fig. 1. Most of the initial ε_{Nd} values are based on the Sm-Nd mineral isochrons~~,~~ and should thus give reliable estimates for the initial isotopic composition of the rocks in question. It should be noted, however, that some data are based only on analyses of whole rocks, and that some points have a relatively large error in age. Nevertheless, the data show a wide range in the initial Nd isotopic composition in the pre-1.9 Ga mafic rocks in the Fennoscandian Shield. Some rocks clearly were derived from a depleted mantle source, whereas others show a major contribution from old enriched lithosphere. It can be questioned whether this is due to crustal contamination, preferably at deep crustal levels, or due to heterogeneity of the subcontinental lithospheric mantle.

The diagram also gives some previously published results from mafic volcanic rocks in Finland. These include the Sm-Nd ages for the Jeesiörova komatiites (2056±25 Ma, *Hanski et al 2001b*), and Vesmajärvi tholeiites (1982±41 Ma, *Hanski & Huhma, in press*), which both were based on analyses of primary igneous pyroxene together with whole rocks. The age for Vesmajärvi metavolcanites is confirmed by a U-Pb zircon age of 2017±5 Ma obtained for several coeval felsic rocks (*Rastas et al., 2001*), which suggests that the initial $^{143}Nd/^{144}Nd$ ratios in the Vesmajärvi pyroxenes and whole rocks were the same (κ_{Nd} = +3.8).

The well-known 2.44 Ga layered intrusions manifest the first major Paleoproterozoic mafic magmatism, which have generally been considered as a sign of break-up of an Archean continent. Our studies have brought new occurrences to this family, i.e. the mafic intrusions of Tsohkoaivi in NW Lapland, Peuratunturi and Koulumaoiva in E Lapland, and several dikes crosscutting the Archean crust. It has also become clear that some mafic dikes show positive initial κ_{Nd} values, which strongly contrast the general picture, i.e. the 2.44 Ga layered intrusions typically have negative initial κ_{Nd} values of –1 to –2 (e.g., Huhma et al. 1990).

Also ultramafic metavolcanic rocks from the Mäntyvaara Formation in Salla, E Lapland provide initial κ_{Nd} values at 2.44 Ga close to -2 (*Hanski & Huhma, in press*). The age is constrained by the crosscutting Onkamonlehto dike providing a minimum age of 2.33 Ga for the volcanic rocks (*Manninen & Huhma, 2001*). These LREE-enriched komatiites resemble the komatiites from the Vetreny belt (*Puchtel et al., 1997*). The low κ_{Nd} values suggest that REE in these rocks were largely derived from lithospheric sources, which were enriched in LREE in late Archean times.

The age of the 2.3 Ga mafic rocks is well constrained in two places, i.e. at Tulisaari (Varpaisjärvi) and Karkuvaara (Pudasjärvi). The initial Nd isotopic composition (κ_{Nd}=+1.8) suggests that the mafic magma had an origin from a depleted mantle without major communication with Archean LREE-enriched material, and is thus very distinct from the bulk of the 2.44 Ga magmatism.

The 2.2 Ga differentiated mafic intrusions of the gabbro-wehrlite association (GWA) from distant places in the shield form a coherent group in the age-κ_{Nd} diagram (Fig. 1). These rocks are characterized by a nearly chondritic initial Nd isotopic composition (κ_{Nd} = +0.6), which was also obtained for a granophyre in the Koli intrusion. This suggests that in situ contamination was not important in the genesis of these rocks, and the chemical composition of GWA suggests an origin from the subcontinental lithospheric mantle.

Many 2.1 Ga mafic rocks are characterized by positive $\kappa_{Nd}(t)$ values, which approach the composition of the (model) depleted mantle. This may be due to major attenuation of the lithosphere, which eventually allowed material from the convective mantle to escape to the surface. This is well exemplified by the Jeesiörova (Sattasvaara) komatiites and Jouttiaapa basalts. In contrast, the ore-bearing 2.06 Ga Keivitsa intrusion in Lapland shows major contribution from old enriched lithosphere. Particularly, the Ni-PGE-bearing ore type from Keivitsa with $\kappa_{Nd}(t)$ of –6.4 shows characteristics typical for late Archean crust (Fig. 1).

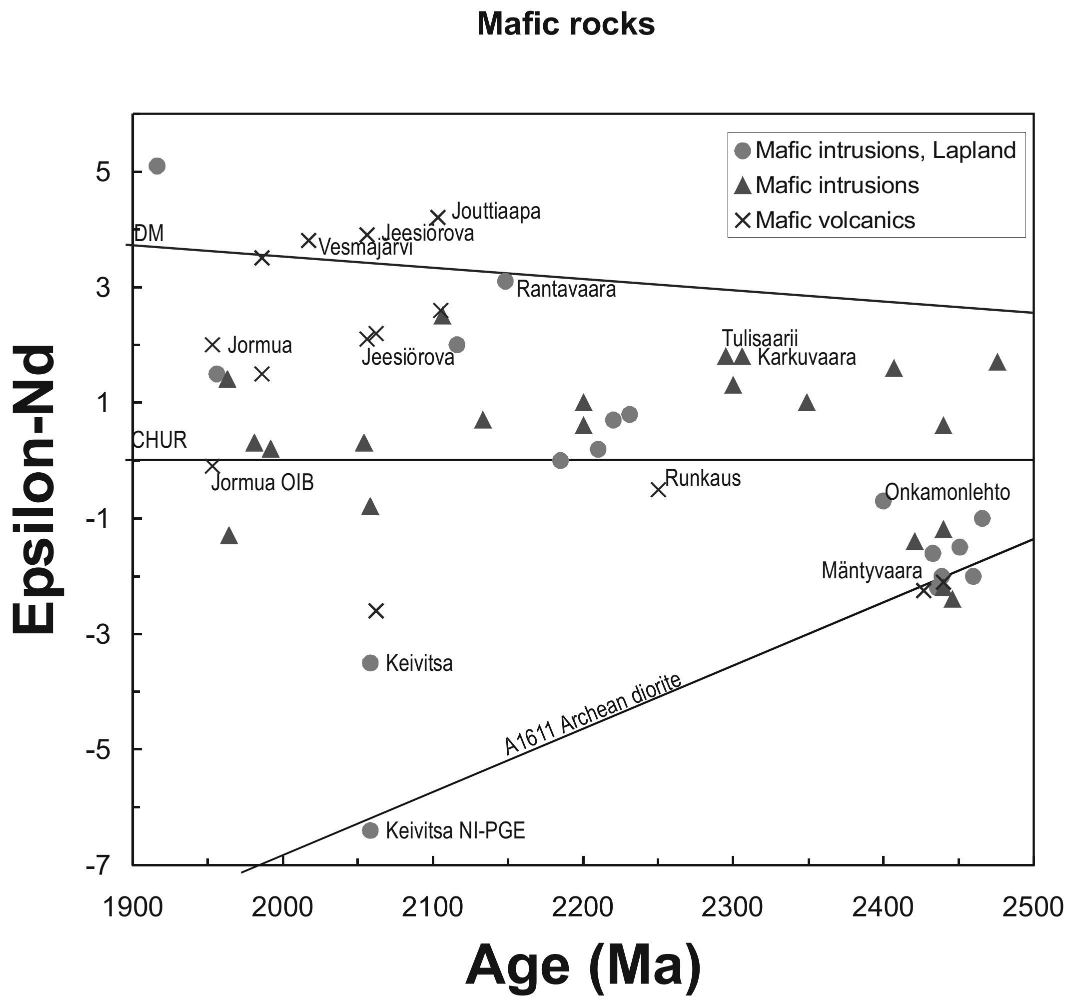

Figure 1. ε_{Nd} vs. age diagram for some pre-1.9 Ga mafic rocks from northern and eastern Finland and adjacent Russia.

3. Remarks

Deep-crustal contamination of ultramafic magma may explain many features observed in mafic rocks, but the isotopic results also show that various mantle sources with distinct isotopic compositions have existed during the Paleoproterozoic. Further examples are provided by high-REE mantle-derived rocks, which show a range of initial ε_{Nd} values from nearly chondritic (e.g. 2.6 Ga Siilinjärvi carbonatite, 2 Ga Jormua OIB, 1.8 Ga lamprophyres) to highly positive (e.g., the ca. 2 Ga Laivajoki and Kortejärvi carbonatites).

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Seismic tomographic images of the crust in southern and central Finland

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Seismic velocity structures of the crust in South Finland are inverted from first P- and S-arrivals of 238 local events. The data set consists of 142 local events recorded at permanent seismic stations in 1992-1993, 26 controlled shots recorded at permanent stations and at the receivers on the DSS lines during the refraction surveys in 1981, 1982, 1991 and 1994 as well as 70 seismic events recorded by the SVEKALAPKO seismic tomography array in 1998-1999. Altogether 7700 P-phase and 6300 S-phase picks are inverted to velocity structures applying the tomography program Jive3d. The target volume is situated between longitude 20°-32-°E, latitude 59°-66°N and depth 0-70 km.

Keywords: Seismic tomography, v_p /v_s –ratio, crust, Finland

1. Introduction

Local tomography inversions are commonly applied to local and regional earthquake data in seismically active regions, or to controlled source experiments with dense net of receivers and with hundreds of sources. In southern and central Finland earthquakes are weak (magnitude < 4) and rarely occurring, and the crustal structure is very complex and exceptionally thick (*Luosto, 1997*). The most prominent geological structure is the fractured suture zone between the Archean and Proterozoic parts crossing the study area in NW-SE direction.

In 1998/1999 a large-scale experiment, the Europrobe/SVEKALAPKO seismic tomography array (*Bock and the SSTWG, 2001*), was carried out by installing a network of 148 stations covering the whole south-central Finland and extending to the Russian Karelia (Fig. 1). The network recorded over 500 local and regional seismic events (earthquakes and explosions) during the nine months long operation period.

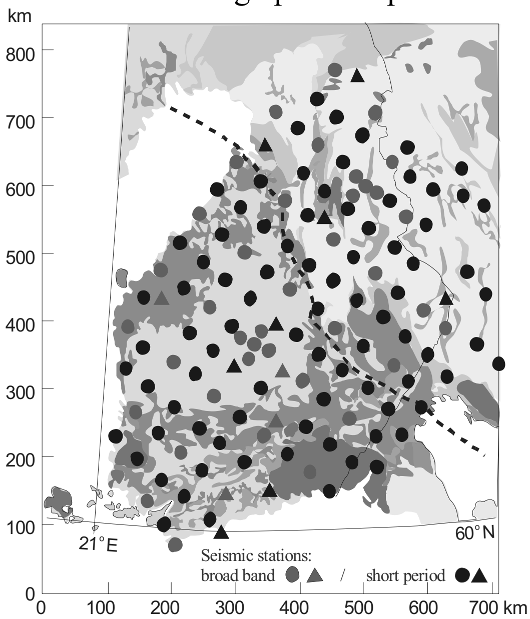

Figure 1. The general geological map of the study area (*Koistinen at al., 2001*) is shown together with the SVEKALAPKO station network.

In order to model 3-D crust velocity structure below the study region, the first arrival traveltimes of P- and S-waves from local explosions and earthquakes recorded by the SVEKALAPKO seismic array together with the first arrivals from explosions, earthquakes and controlled source shots recorded by permanent seismic stations and the DSS receivers (Fig. 2), are inverted. The program package Jive3D by *Hobro* (*1999*) in which a linearized, regularised model space is inverted by optimizing the model fit with travel time data to result in smoothed velocity structures in agreement with the data density and quality, is applied.

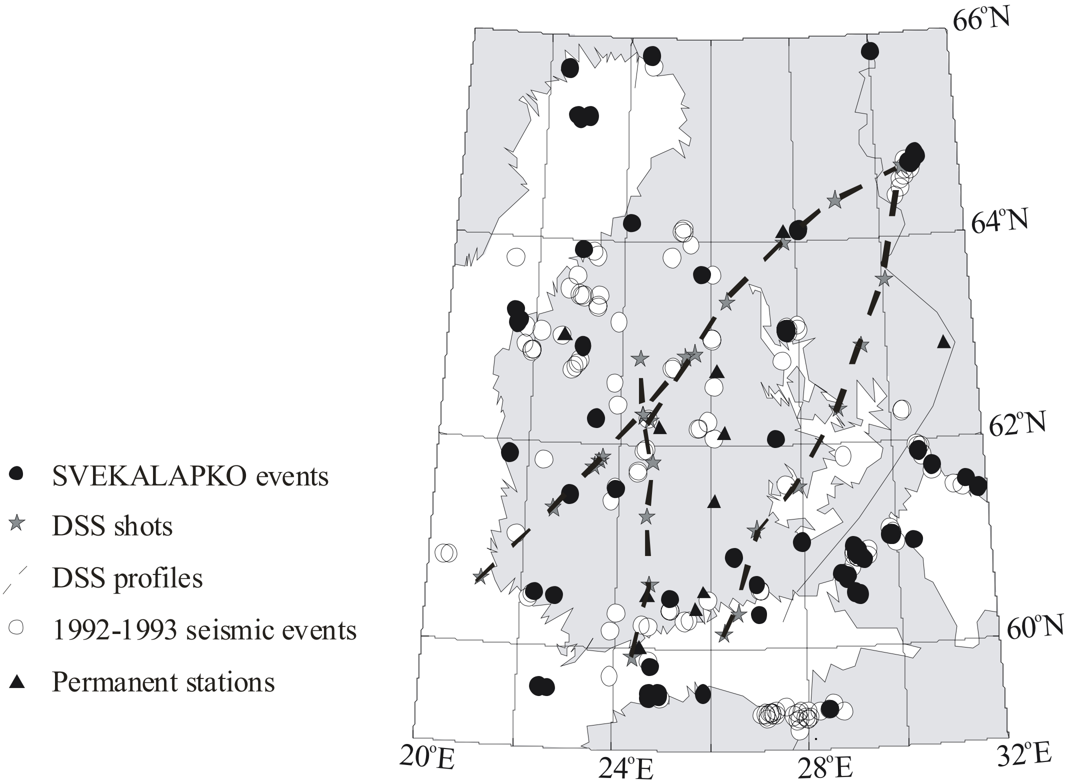

Figure 2. Location of the seismic sources, DSS profiles and permanent stations in the study area.

2.Velocity model

The model is parameterized as a regular grid of evenly spaced velocity and depth nodes, and the velocity fields are computed using quadratic B-splines. Jive3d uses two-point ray tracing with shooting method, linearized iterative approach, regularised least squares inversion and conjugate gradient method for optimization. The convergence is controlled by the step-by-step decrease in regularisation strength.

The P- and S-velocities are inverted separately. Different initial models such as laterally homogeneous models with vertical velocity gradients and a three dimensional crust model derived from earlier refraction and reflection studies (*Sandoval et al, 2003*) are tested. The inverted velocity patterns have proved to be independent of the start model.

3. Tomographic inversion

After inverting the first output models the arrivals with residuals over 2 seconds are rejected. The non-controlled sources are relocated with a semi-final model by grid search method, and the final models are inverted with the new locations. The resolution of the inversion is carefully studied with checkerboard test by using different sizes of cells. The

v_p/v_s-ratio distribution (Figs. 3 & 4) is limited by the resolution of the S-waves, which is worse than the P-wave resolution. The results are very sensitive to data density and quality, and most of the inaccuracies originate from the uncertainties of quarry blast and earthquake locations or from the inaccuracies of the phase onset times.

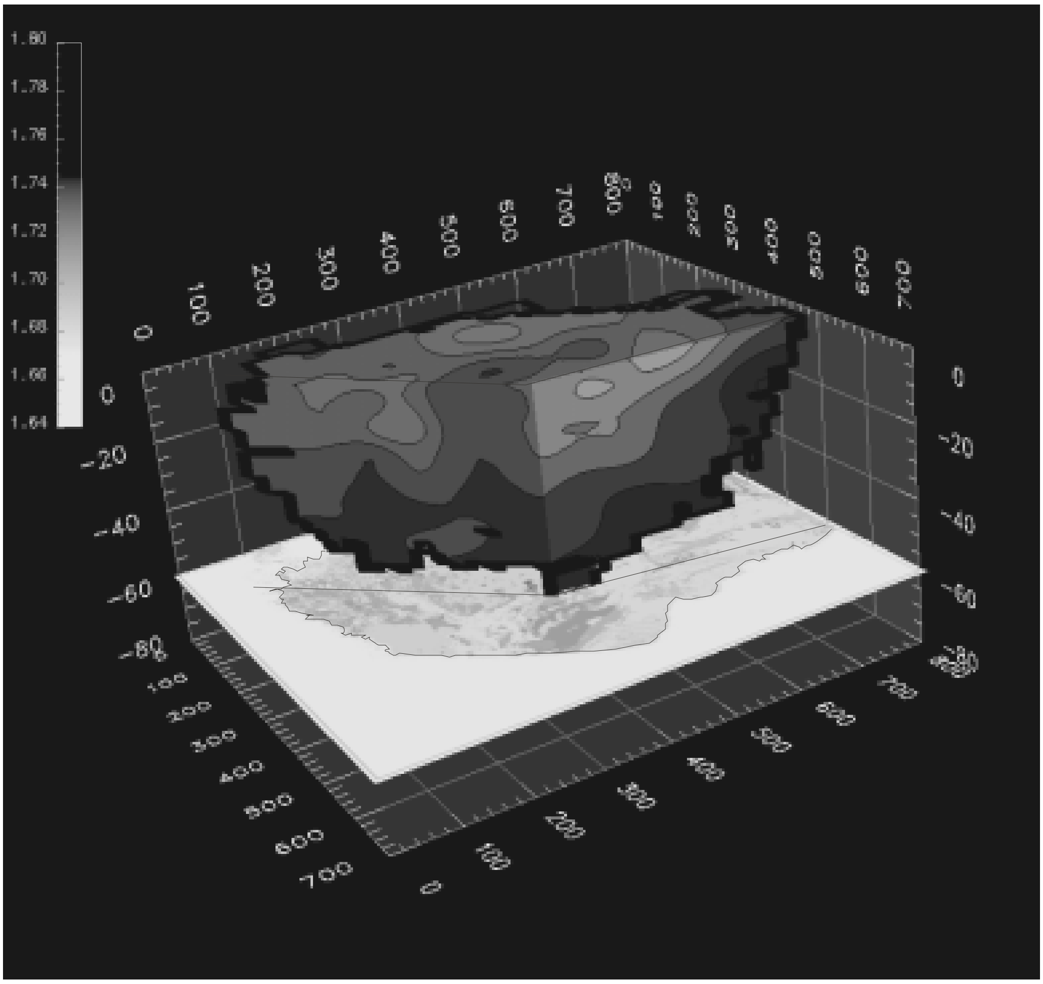

Figure 3. A three-dimensional section of the v_p/v_s-ratio distribution is presented for southern and central Finland looking towards northwest. The intersecting lines are marked on the geological map. The image is created by a 3-D graphical user interface built of modules from the OpenDX visualization software package (*Thompson et al., 2004*).

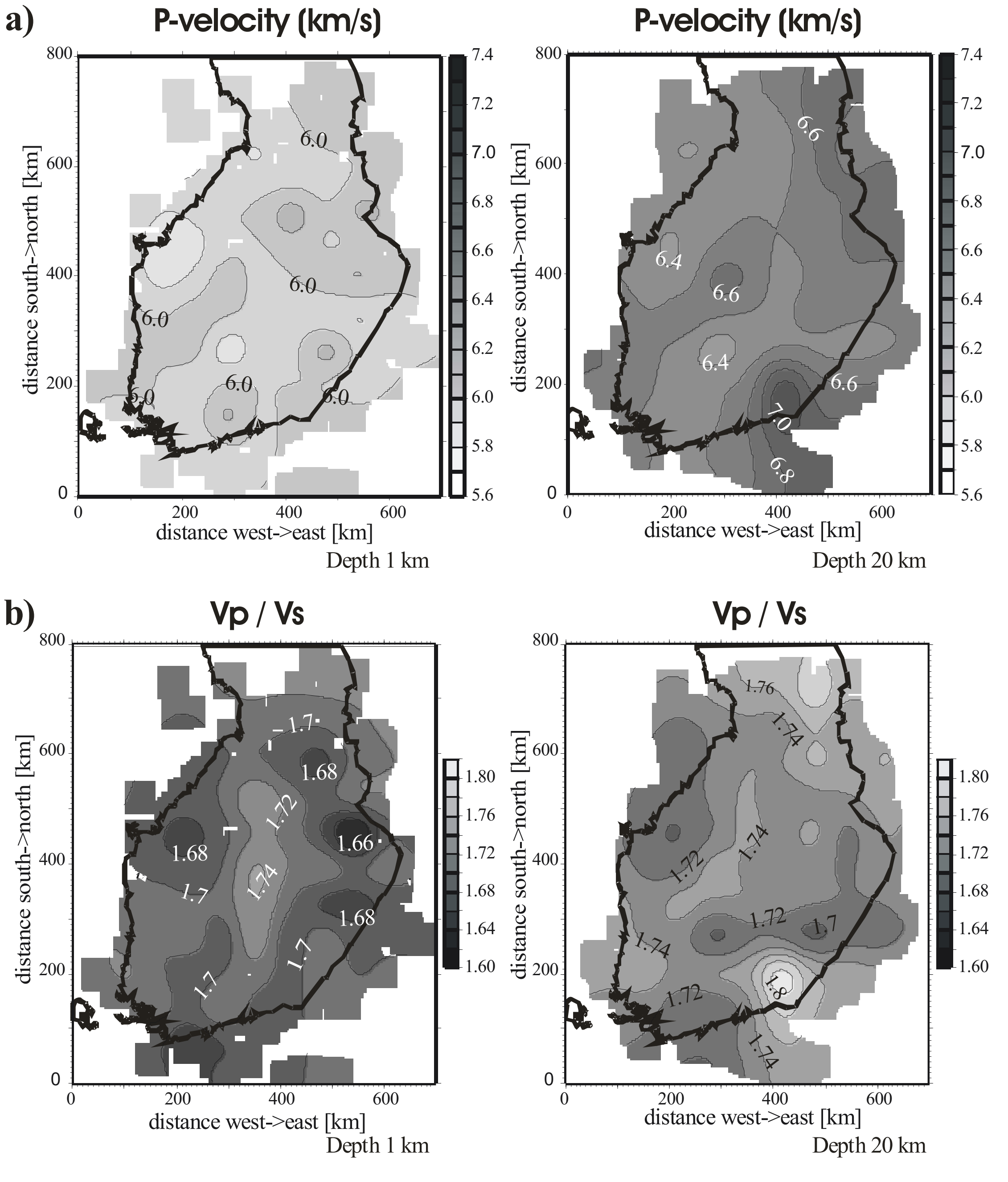

Figure 4. Results of the tomographic inversion at depths of 1 and 20 km. a) P-velocity and b) v_p/v_s-ratio distribution. The white colour masks out the regions with no rays.

4. Conclusions

The final velocity models are smooth and in accordance with the sparse ray coverage over a large target volume. The horizontal resolution of the velocity structure is at least 60 km in most parts of the models. Tomographic velocity models agree with the 2D-ray tracing results of the DSS studies.

The v_p/v_s-ratio varies usually between 1.66-1.76 in the upper crust and between 1.72-1.82 in the lower crust. In the upper 10 km, the Archean and Paleoproterozoic domains are outlined by velocity differences in both P-velocity and v_p/v_s-ratio maps (Figs. 3 & 4). Higher velocities and lower v_p/v_s-ratios north-east of the isotope boundary at the surface (broken line in black in Fig. 1) are associated with the Archean lithologies.

The Bothnian and Southern Finland Schist Belts have lower v_p/v_s-ratio than the Central Finland Granitoid Complex. Rapakivi granites of the upper crust are associated with high v_p/v_s-ratio rocks in the middle and lower crust.

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A research project: An integrated 3-D geophysical model of the lithosphere in the central Fennoscandian Shield

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We present a new project that aims to integrate all the recent geophysical data available for central Fennoscandian Shield, i.e., seismic, potential field and electromagnetic data, and to create a 3-D geophysical model of the lithosphere that includes distribution of P- and S-wave velocity, density and electrical conductivity and give the explanation to this model in terms of rheology, pressure-temperature conditions and compositional variations.

Keywords: Fennoscandian Shield, crust, upper mantle, joint inversion, seismic tomography, potential fields, electromagnetic deep sounding

The research project *An integrated 3-D geophysical model of the lithosphere in the central Fennoscandian Shield* is a part and simultaneously a continuation of several on-going projects carried out by the Department of Geophysics of University of Oulu and Oulu Unit of Sodankylä Geophysical Observatory of University of Oulu in collaboration with the Geological Survey of Finland and other research departments from Europe and Russia.

Two of the projects, the *SVEKALAPKO Deep Seismic Tomography project* and *BEAR, or Baltic Electromagnetic Array Research*, are two important parts of multidisciplinary international EUROPROBE/SVEKALAPKO research, aiming at studying the evolution and structure of the lithosphere below Finland. The third project, called *"3-D crustal model of the Finnish part of the SVEKALAPKO research area"* (3-DCM project) aims at joint interpretation of the SVEKALAPKO seismic data and potential field data in collaboration with the Geological Survey of Finland. A new portion of high-quality geophysical data was provided by two most recent projects. *The FInnish REflection experiment (FIRE)* started in 2001 and measured about 2100 km of reflection seismic profiles across the main geological boundaries in Finland (*Kukkonen et al., 2002*). *The MT-FIRE project* started in 2003 and aims to collect high-density AMT-MT data (period range 300 Hz – 10000 s) at two profiles collocated with the reflection seismic profiles FIRE-1 and FIRE-3 and along one profile between the two seismic profiles.

Although the EUROPROBE program of the European Science Foundation was officially finished in 2001, the interpretation of the data collected within its different projects and preparation of publication continues at least till the end of 2005. As a part of this final stage, the SVEKALAPKO Seismic Tomography Working Group and the BEAR Working Group are now continuing the data interpretation and integration of results of different methods (Hjelt and Daly, 2002). The EUROPROBE/SVEKALAPKO project was funded by the Academy of Finland till the end of 2002, while funding of the 3-DCM project continues until December 2004.

The recent results of the SVEKALAPKO Deep Seismic Tomography project (Alinaghi et al., 2003, Bruneton et al., 2002, Sandoval et al., 2002, 2003, Yliniemi et al., 2004) and the BEAR project (Engels et al., 2002) contain a lot of completely new information about the structure of the upper mantle below the SVEKALAPKO research area. Proper explanation of these results in terms of petrophysics, rheology and pressure- temperature conditions requires further modelling and integration with other geophysical methods that can provide self-consistent, integrated geophysical model of the lithosphere in the central Fennoscandian Shield. Joint interpretation of global gravity and tomography data by *Deschamps et al*. (*2001, 2002*) demonstrated that such an approach allows to give explanation of tomographic models in terms of thermal and compositional anomalies (iron content). Joint interpretation of SVEKALAPKO seismic data and gravity data has been also successfully applied in the 3-DCM project (*Kozlovskaya et al., 2004*).

That is why the main target of the proposed research is to integrate all the recent geophysical data available for central Fennoscandian Shield, i.e., seismic, potential field and electromagnetic data, and to create a 3-D geophysical model of the lithosphere that includes distribution of P- and S-wave velocity, density and electrical conductivity and give the explanation to this model in terms of rheology, pressure-temperature conditions and compositional variations. The model will provide a better image and understanding of processes that resulted in the present-day structure of the Archaean and Proterozioc lithosphere in the Fennoscandian Shield.

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BIFROST: Continuous GPS measurements of the three dimensional deformation in Fennoscandia

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Ongoing BIFROST (Baseline Inferences for Fennoscandian Rebound Observations, Sealevel, and Tectonics) project uses a network of permanent GPS stations in Finland and Sweden to measure crustal deformation due to glacial isostatic adjustment (GIA). Project was initiated in 1993. In this paper we concentrate on the geodetic results of the project. We show series of 3D velocities that may be used as a new set of observables for GIA adjustment.

Keywords: vertical motion, space geodesy, glacial isostacy

1. Introduction

Networks of permanent geodetic GPS (Global Positioning System) stations are providing a new method to determine present day crustal deformations. Modern space geodetic techniques allow us to measure both horizontal and vertical motions simultaneously in a well-defined reference frame to observe glacial isostatic adjustment (GIA). Over last 10 years the accuracy of the GPS determination has increased significantly. This is mainly due to advantages in GPS satellite constellation, GPS receiver design and processing techniques. The horizontal position estimates can be nowadays achieved at a few millimetre level and vertical rates some 2 times less accurately. Baseline Inferences for Fennoscandian Rebound Observations, Sea level, and Tectonics (BIFROST) is a project that was initiated in 1993 taking advantage of the tens of permanent GPS stations both in Finland and Sweden. The goal of the BIFROST is to directly measure the present day crustal deformation in Fennoscandia and provide a new GIA observable for determination of the Earth structure and Fennoscandian ice history. In this paper we present the results from the time series of nearly 2500 daily network solutions. (*Johansson et al., 2002*). BIFROST project has already obtained significant results in this regard (*Johansson et al. 2002*), (*Milne et al., 2001*), (*Milne et al. 2004*), (*Scherneck et al., 2002*).

2. The BIFROST GPS network

The BIFROST GPS network consists of permanent networks in Finland and Sweden (Figure 1). In addition, also some of the IGS (International GPS Service) stations outside the area are included into analysis. A full documentation of the stations, equipments and their histories are given in *Johansson et al., 2002*.

The Swedish nationwide SWEPOS™ GPS network of 21 stations was established in 1993 and became fully operational in 1995. The standard monument for the SWEPOS™ station is a 3 m heated concrete pillar. The pillars are built on the bedrock. Onsala belongs both SWEPOS and IGS network. At Kiruna SWEPOS and IGS stations are separated by 5 km. The stability of the pillars is annually controlled by the theodolite measurements. The GPS antenna is replaced by a theodolite and horizontal and vertical angles are measured to a nearby network of steel pins. All SWEPOS™ stations are equipped with two or more GPS receivers which are connected to a single Dorne Margolin type antenna. (*Johansson et al., 2002*), (*Scherneck et al., 2002*).

The Finnish permanent GPS network FinnRef® consists of 12 GPS stations. The network became fully operational between 1994 and 1996. Metsähovi station belongs both FinnRef and IGS network. Three different types of antenna platforms are used. A standard type is a 2.5 m heigh steel grid mast. It is estimated that the thermal expansion effect is less than 1 mm in a yearly cycle. The thermal expansion can also be controlled with temperature measurements. On three of the stations we have concrete pillars and on two stations we use higher masts. In the latter case the thermal expansion is compensated using an invar steel suspension system. The stability of the masts is also controlled with tachymeter measurements. In FinnRef stations the antenna is not removed during the tahymeter measurements but the antenna has a permanent pin under it on which the angles are measured from the auxiliary benchmarks around the pillar. The control measurements are performed every second or third year. (*Johansson et al., 2002*) (*Koivula et al., 1997*), (*Koivula et al., 1999*).

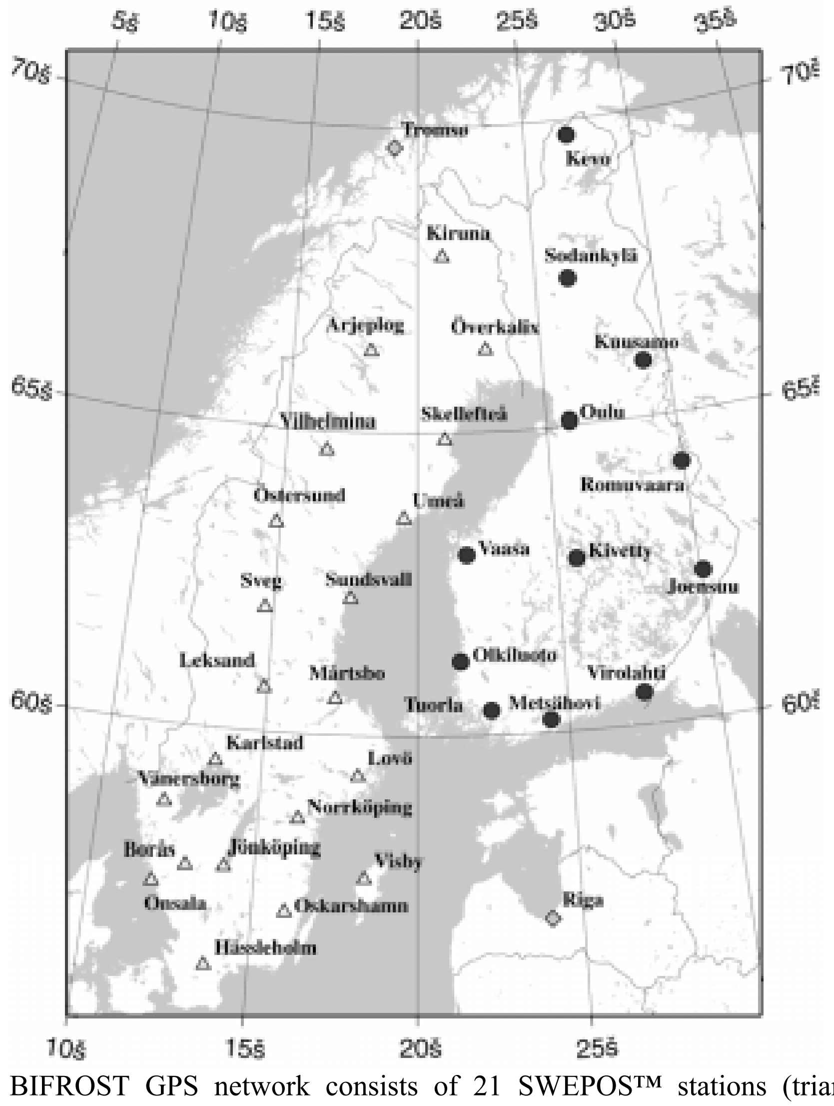

Figure 1. (a) BIFROST GPS network consists of 21 SWEPOS™ stations (triangles) and 12 FinnRef® stations (circles). In the processing the data from IGS (International GPS Service) is also included (diamond).

Figure 2. Photograph of the FinnRef station at Romuvaara where the antenna platform is a concrete pillar. A Dorne Margolin type antenna is covered by a conical radome. (Photo: H. Koivula)

3. Data analysis

All the data from BIFROST GPS stations are collected in the data archive at Onsala. For the processing the dual frequency phase and code observations are used. The data are processed with GIPSY software which was developed at Jet Propulsion Laboratory (*Webb and Zumberge, 1993*). Alternative solutions with Bernese software (*Hugentobler et al., 2001*) have been computed at Onsala and at the Finnish Geodetic Institute.

First the original GPS data is re-sampled into 300 s observing interval. The data are processed in 30-hour sessions. For each session we estimate the receiver clock correction, three site coordinates, atmospheric zenith parameters and ambiguity parameters. The satellite orbits were strongly constrained to the precise orbit values distributed by the IGS (International GPS Service). The data below 15º were not used in the processing. Corrections for ocean loading and solid Earth tides are included into the model. The data are processed using non-fiducial technique where all stations have weak a priori constraints. Traditional fiducial technique where a network is surrounded by known stations which are kept fixed was rejected since we thought that the stability and the continuity of BIFROST network is better than the IGS network in general. The geodetic results shown in chapter 3 are processed without fixing the ambiquities. Tests show that 85-100 percent of ambiquities could be solved even over these long baselines. Independent tests in smaller data sets show that fixing ambiquities decreases the formal errors 20-30%. In the future the computer power and the Software are able to perform also ambiguity fixing in a reasonable time. Finally all the geodetic positions obtained by GIPSY are transformed into a common terrestrial reference frame. (*Johansson et al., 2002*) (*Scherneck et al., 2002*)

4. Geodetic Results

Our standard geodetic solution includes a mean value, a constant rate, an admittance parameter for atmospheric loading and periodic terms with different frequencies. Also a bias parameter was included on times that were expected to influence on the rate estimates. These kind of known interruptions are changes or removals of antennas or antenna radomes etc. Numerical results are given in Johansson et al., 2002. In figure 3 we show an example of time series of three GPS stations. Metsähovi is a FinnRef station, Mårtsbo and Norrköping are SWEPOS™ stations. It is obvious that the time series have certain jumps. These jumps are connected to changes in antenna radomes or antennas or even to heavy snowstorms and introduced as bias into the rate estimation.

We find that the maximum uplift rate is ~10mm/yr. This agrees with the prediction by Lambeck (Lambeck 1998a, 1998b) better than 1 mm/yr. The results have been also compared to an independent GPS solutions and to the ones from the sea level and repeated precise levellings (*Johansson et al., 2002*) (*Mäkinen et al., 2003*). Generally, one can conclude, that the agreement with different methods are good, although some disagreements were discovered (*Mäkinen et al., 2003*). Some of these can be addressed to the data problems in GPS time series, mainly due to the snow accumulation on some Northenmost stations. Annual variations are also clearly visible (*Poutanen et al., 2004*). Also, horisontal movements are visible in data sets. These are within the model predicton of the GIA (*Milne et al., 2001*).

5. Conclusion

BIFROST project has shown a power of long GPS time series in crustal deformation studies. A new set of observables has been created for GIA adjustment. The geodetic results of the BIFROST project provide a dense three dimensional determination of ongoing crustal deformations in Fennoscandia. Numerical values are published in (*Johansson et al., 2002*). These values has been used and tested in various occasions. The results have already been used for predictions for Fennoscandian ice histories and estimating the bounds for lower and upper mantle viscocities (*Milne et al., 2001*), (*Milne et al. 2004*).

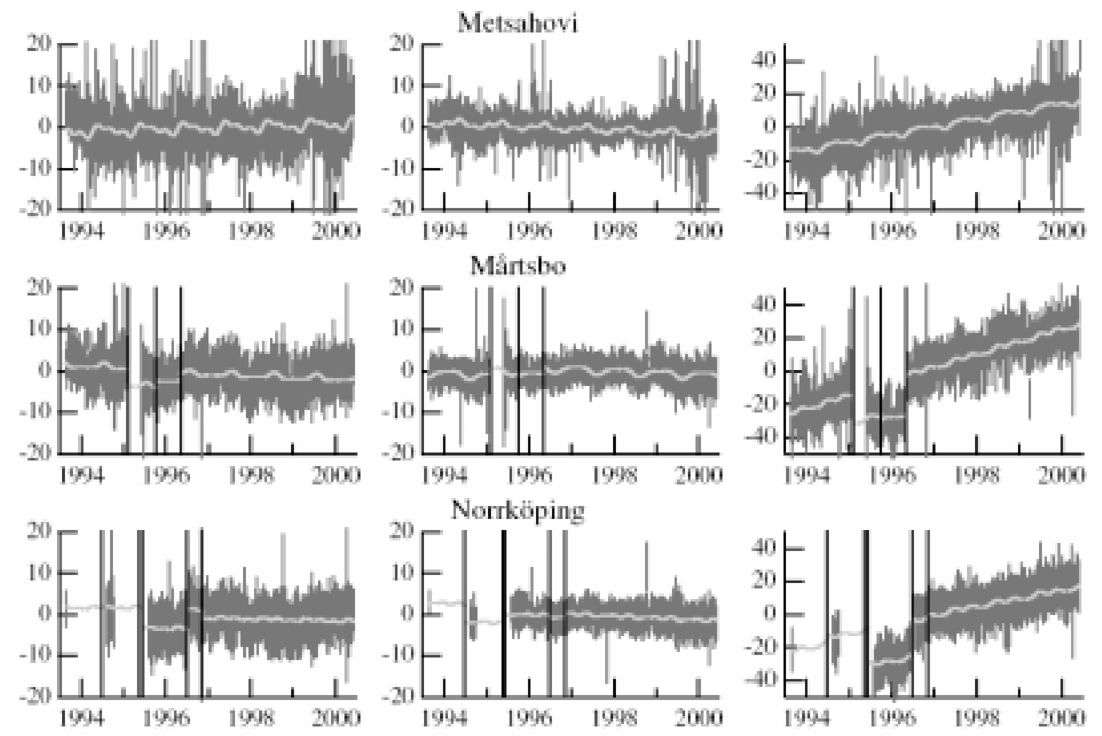

Figure 3. Three examples of the BIFROST time series. Pictures show the differences in mm from the mean of the series. East component is at left, north component in the middle and the radial (height) component at the right. Light line indicates a model fit into the time series and vertical lines show places where offsets were introduced to time series. (*Johansson et al., 2002*)

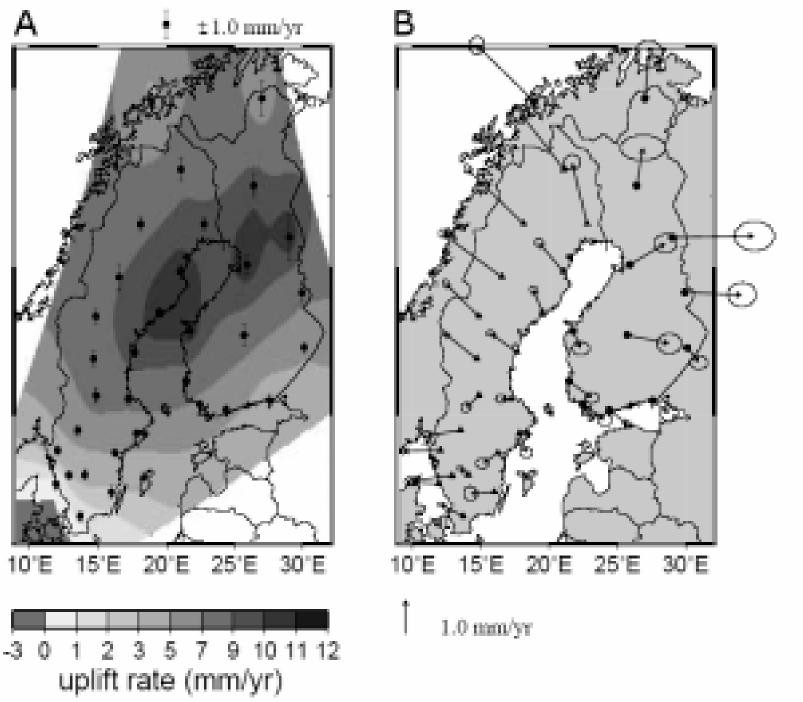

Figure 4. The vertical and horizontal velocity fields according to BIFROST GPS processing. The GPS stations are shown as black circles. (a) The contour lines of a present day vertical velocities. Site specific 1ω uncertainty is shown on every GPS station. The scale is shown at the top. (b) The present day horizontal velocities at each GPS station. 1ω error ellipses are in a scale shown at the bottom. (*Milne et al., 2004*).

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World Digital Magnetic Anomaly Map - presenting lithospheric contribution to the Earth's total magnetic field

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Magnetic field of the Earth consists of two major, slowly changing components, namely the Core (Main) field and Lithospheric (Anomaly) Field. These components can be measured together only, and their separation without knowing either of the sources is an nonunique problem. The anomaly field is caused by magnetization of the upper lithosphere and hence reflects its geological properties. In lithospheric studies these are interpreted from the estimated anomalies and other geophysical-geodetic information. For years the International Association of Geomagnetism and Aeronomy (IAGA) has encouraged making a global magnetic anomaly map for science and education. Its task group for World Digital Magnetic Anomaly Map (WDMAM) aims to complete the first edition of the map as an international, geophysical-geological joint effort until 2007.

Key words: Magnetic anomaly, Magnetization, Lithosphere

1. 1979 and Boulder 1995, IUGG/IAGA presented resolutions to compile nationa. At Canberra l magnetic anomaly maps 1:2.5 million for further compilation of maps of the continents and finally making the anomaly map of the World plus releasing the data base beyond the map (*Reeves et al. 1998*). Working group for Magnetic Anomalies, Land and Sea (lately WG V9) supervised the task until IUGG/IAGA at Sapporo 2003, where a Task Force (WDMAM) was established to continue the work. During IGC32 at Florence 2004, co-operation between IUGG/IAGA and IUGS/CGMW was initiated for map completition. *(Korhonen et al. 2004*)

2. The global magnetic anomaly map aims to display such component of the magnetic field that is caused by the ferrimagnetic uppermost part of the Lithosphere. The map will be smoothed as if the component were observed at an altitude of a few km above the Earth's surface. It will be suitable to study overall crustal structure, make regional and global crustal models, and serve as background information for more detailed geological and geophysical studies.

3. The map will be based mainly on airborne and seaborne magnetic surveys (*e.g. Finn et al. 2001*). Satellite magnetic data will be used for such parts where near ground information is poor or missing (e.g. *von Frese et al. 2002*). Ground measurements will be used from magnetic observatories and areas where extensive ground surveys provide principal regional coverage (e.g. *Korhonen et al 2002*). At the moment open questions still remain to be solved, including anomaly calculation schemes and rights of data access and database distribution. The WDMAM plans to arrange a scientific session of magnetic anomaly compilation and show a tentative global anomaly map in the next IAGA meeting in Toulouse 2005. Further, the first edition of the map would be published and the database released in IUGG meeting in Perugia 2007.

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Introduction to the WDMAM:
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Antarctic:
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The processes forming the Palaeoproterozoic Svecofennian

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This paper focuses on the processes associated with the formation of the Svecofennian. It is suggested that the Svecofennian domain comprises of at least five orogens each having compressional and extensional stages. The complexity arises from interacting processes, small plates and different tectonic environments existing on opposing margins of a plate at a given time. In addition to the plate driving forces also gravitational forces were important during Palaeoproterozoic crustal and lithospheric growth.

Keywords: Svecofennian, Fennoscandia, lithosphere, orogeny

1. Introduction

Plate boundary forces create variable tectonic environments that overlap in time and space. The Palaeoproterozoic Svecofennian domain of Fennoscandia (Figure 1) is comparable in size and complexity to Cenozoic Central Europe, and to present Indonesian archipelago where several processes operate simultaneously and where the change from one tectonic setting to another is rather abrupt

An orogeny is by definition a process of creation of mountain belts by tectonic activity *(Bates and Jackson, 1995*). Thickening of the lithosphere during collision induce gravitational and mechanical instabilities that tend to be stabilized instantaneously (Figure 2.). When the stabilization processes are included the orogenic cycle is characterized by thickening of the crust, thermal maturation, partial melting and syn- to post-convergence gravitational collapse *(Vanderhaeghe and Teyssier, 2001)*.

Palaeoproterozoic orogenies have been interpreted to be formed in a compressional regime implying semicontinuous convergence over a long time, and as a consequence the stages of these orogenies have been called synorogenic, lateorogenic and postorogenic. One example of the proposed long-lasting orogeny is the Svecofennian Orogeny (100 Ma; *Gáal and Gorbatschev, 1987*). If it is assumed that the forces acting on the plates were in the Precambrian similar to those acting on the plates today, the long-lasting orogenies with only collisional phases are impossible and other more complicated explanations with both collisional and extensional stages must be looked for.

2. Geological background

At the southern rim of northern Svecofennian (Northern Svecofennian subprovince of *Gáal and Gorbatschev, 1987*; Figure 1) the oldest lithological units comprise few older volcanic rocks and granites (>1.95 Ga; Knaften (K)) south of the Skellefte district (SD) and 1.92 Ga old tonalitic rocks interlayered with volcanic and turbiditic rocks in the Savo Belt (SB). The area was later intruded by synkinematic granitoids between 1.89 Ga and 1.88 Ga and by post-kinematic pyroxene-bearing granitoids at 1.885 Ga. The Skellefte district (SD; 1.89-1.88 Ga) in the northern Svecofennian of Sweden is composed of two groups of calc-alkaline metavolcanic and metasedimentary rocks intruded by a variety of granites.

In the eastern part of the Central Svecofennian (Central Svecofennian subprovince of *Gáal and Gorbatschev, 1987*) the Central Finland Granitoid Complex (CFGC) comprising mainly calc-alkaline I-type granitoids (1.88-1.89 Ga) with minor amounts of 1.87 Ga postkinematic pyroxene-bearing granitoids. The complex is surrounded by migmatites with tonalite leucosome that were formed from immature psammites 1.89-1.88 Ga ago. The Bothnian Basin (BB) in the western part of the Central Svecofennian is composed of psammitic metagreywackes interbedded with black shales and minor mafic volcanic rocks as well as 1.89-1.87 Ga old calc-alkaline granitoids.

The Southern Svecofennian (South Svecofennian subprovince of *Gáal and Gorbatschev, 1987*) includes the 1.90-1.89 Ga Bergslagen area (BA) and Uusimaa Belt (UB), partly formed in an intra-arc basin of a mature continental arc. Arc-type volcanic rocks are also found in the Häme belt (HB in Fig. 2).

In central and northern Sweden, granites with similar age distribution (1.84-1.82 Ga) occur in the Bothnian Basin and in the Skellefte district. The latest major magmatic episodes are expressed as granitoid and volcanic rocks in the Transscandinavian Igneous Belt (TIB) (1.80-1.78 Ga) in Sweden.

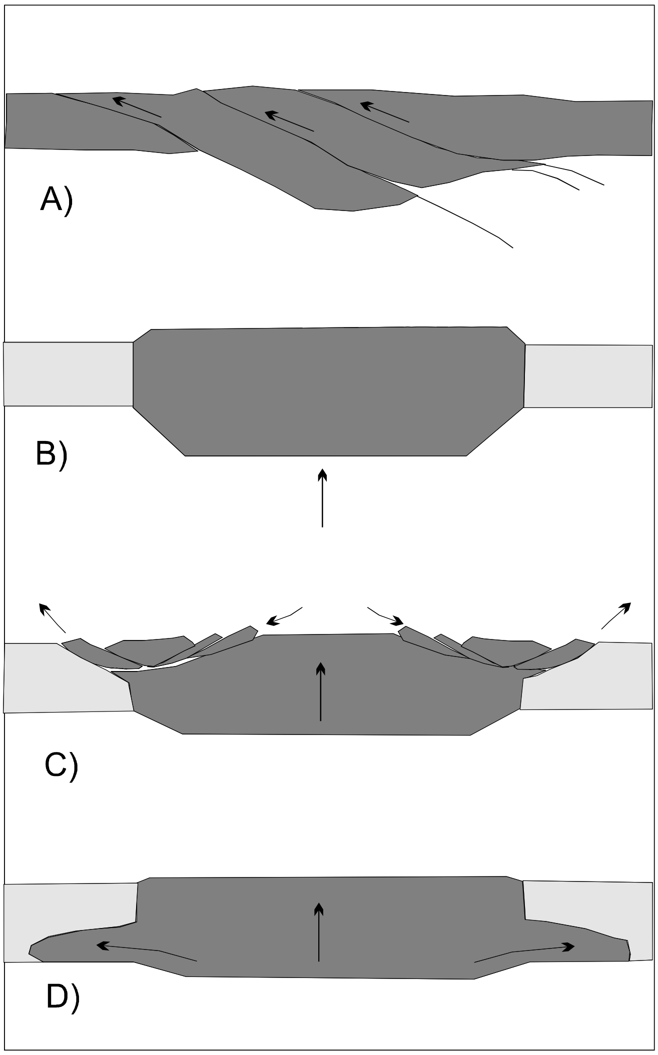

Figure 1. Orogenic processes forming the continental crust after *Coward, 1994; Rey et al., 2001*). a) thickening by stacking of continental slivers, b) unstable thickened crust, c) gravitational collapse by upper crustal deformation only d) gravitational collapse by lower crustal flow only (blind collapse).

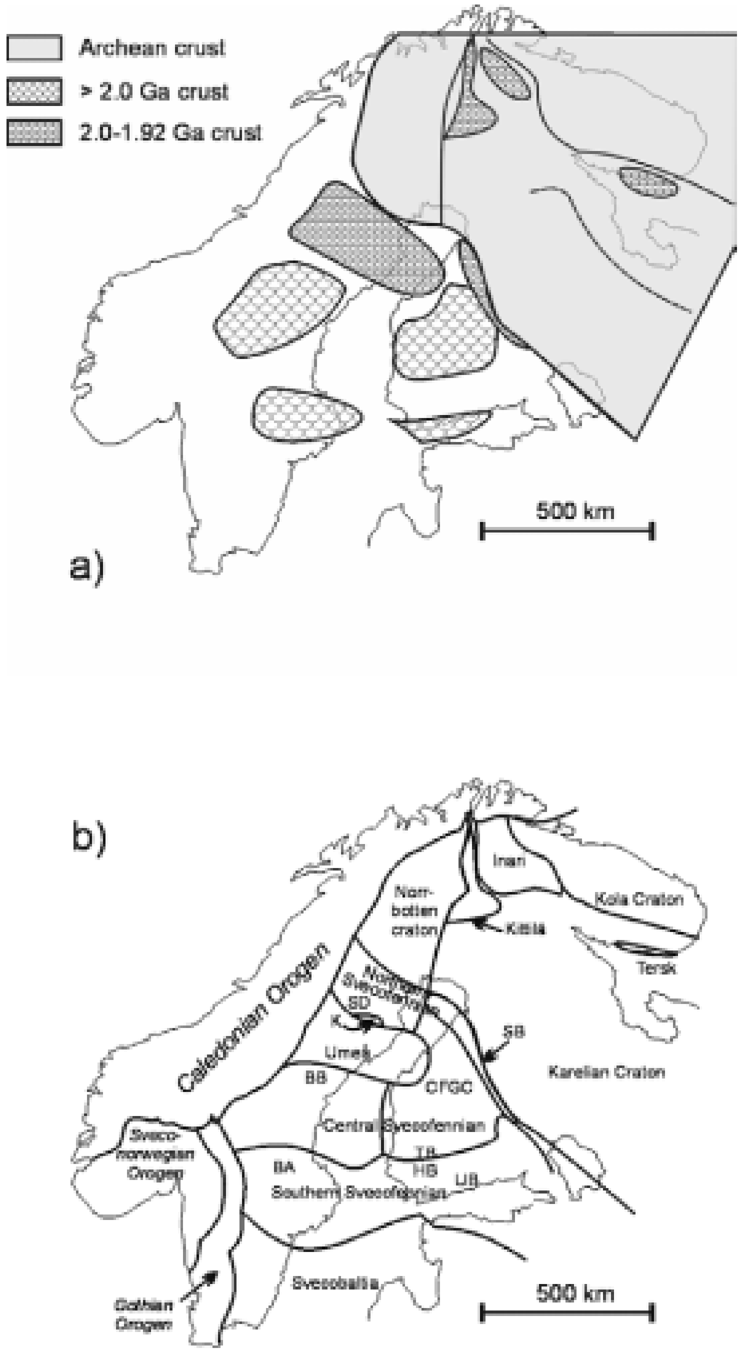

Figure 2. Distribution of microcontinental nuclei and terrane boundaries in the Fennoscandian Shield *(Lahtinen et al., 2004)*.
a)Hidden and exposed pre-1.92 Ga old crustal units of the Fennoscandian Shield.
b) Major Palaeoproterozoic terranes of the Fennoscandian Shield.
Abbreviations: BA – Bergslagen area; BB – Bothnian Basin; CFGC - Central Finland Granitoid Complex; HB – Häme Belt; K – Knaften; SB – Savo Belt; SD – Skellefte district; TB – Tampere Belt; UB – Uusimaa Belt.

3. Tectonic model

In the following the processes operating in the major orogenic stages of the plate tectonic model by *Lahtinen et al. (2004)* are outlined. The model is divided into five stages: 1) continental rifting, 2) micro-continental accretion, 3) extension of the accreted crust, 4) continent-continent collision and 5) gravitational collapse stages.

After the destruction of the Archean continental plate at 2.1 Ga new juvenile arcs were initiated. Arcs of different evolutionary stages started to assemble and to form microcontinents. By 1.89 Ga several small plates carrying continental fragments had accreted to the Karelian continent. Oblique collision caused some of the shear zones at or close to the terrane boundaries. The microcontinental collision stage resulted in thickened lithosphere in the core area.

The newly formed continental interior was thermally and gravitationally unstable, and as soon as active compression shifted towards the margins extension took over in the interior. A collapse took place at 1.87 Ga when the postkinematic pyroxene bearing granitoids intruded in the the Central Finland Granitoid Complex. The collapse spread towards the orogenic fronts where its effects were more profound. This means that while the orogen was being destroyed in the central parts subduction was still in operation at the margins. At the end of the microcontinent accretion stage a larger but unstable continental plate (Fennoscandia) had developed.

By 1.80 Ga lithospheric convergence had brought Fennoscandia in contact with Laurentia in the northeast, Amazonia in the west, Sarmatia in the southeast and an unknown continent in the southwest. The net result was that several continent-continent collisions were localized around the Fennoscandian nucleus.

After the continent-continent collisions Fennoscandia underwent a major stabilization period with gravitational collapses, thermal resetting and late tectono-magmatic episodes. This period is characterized by rapid uplift, voluminous granitoid magmatism and pegmatite intrusions around the Svecofennian nucleus.

4. Discussion

In the model, the Northern and Central Svecofenninan Subprovinces of *Gaal and Gorbatshev (1987)* were formed during the Lapland-Savo and Fennian orogenies. The northern Svecofennian comprises of more juvenile island arc material whereas the Central Svecofennian is cored by older continental crustal blocks (the Keitele and Bothnia microcontinents). The Southern Svecofennian Subprovince was initiated in the collision between the Bergslagen microcontinent and the newly formed Lapland-Savo orogenic belt. The Southern Svecofennian was, however, heavily reworked during the Svecobaltic orogeny induced by the collision of the Sarmatian crustal segment (*Gorbatschev and Bogdanova 1993*) with the Fennoscandian crustal segment. In the model the Southern Svecofennian Subprovince is divided into older Southern Svecofennian and younger Svecobaltia terranes. Another new terrane, the Umeå allochton, is also separated from the northern and central Svecofennian. It resulted from the collision between Amazonia and Fennoscandia at 1.8x Ga (Nordic orogeny). Although there is considerable overlap in the terrane distributions, the evolutionary models differ markedly. In the model of *Gáal and Gorbatshev (1987*) the Svecofennian was formed during a long semi-continuous orogeny, which is here divided into several separate orogenic cycles with extensional collapse stages following intimately collisions.

The youngest zircons from mantle and lower crustal xenolith rocks in eastern Finland verify mantle activity up to 1.8 Ga (*Peltonen and Mänttäri, 2001*). These ages are interpreted to mark gravitational collapse that took place between 1.80 and 1.75 Ga. Large-scale collapse was possible only after all the long-term compressions at the margins had ceased. The collapse was a combination of several smaller collapses and perhaps of the lithospheric delamination after the Nordic orogeny. However, the small volume and lithospheric origin of mantle magmatism in the central parts of Svecofennia does not favour asthenospheric upwelling in the area.

5. Conclusions

It is concluded that the orogenies even in the Palaeoproterozoic Fennoscandia were short and distinct and that they included alternating collisional and extensional stages. The complexity arises from interacting processes, small plates and different tectonic environments existing on opposing margins of a plate at a given time. The duration of the Proterozoic orogenies approaches that of the Phanerozoic ones.

It is concluded that the "Svecofennian orogeny" comprises of at least five orogenic periods each having compressional and extensional stages. In addition to the plate driving forces also gravitational forces were important during Palaeoproterozoic crustal and lithospheric growth. Gravitational collapse must have occurred with or without lithospheric delamination.

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Stabilization processes of Precambrian continental crust - an integrated geophysical-petrological study of the Central Finland Granitoid Complex

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The origin of the Central Finland Granitoid Complex is studied in an interdisclipinary study using both geophysical and geological data. The project focuses on crustal processes affecting thickened continental lithosphere, structures related to the stabilization of the crust, and formation of large igneous complexes.

***Keywords*:** Central Finland, granitoid, reflection seismic

New continental crust and lithosphere are formed in island arcs and ocean plateaus and can be either preserved or destroyed in the following continental collisions. The collisions induce gravitational and thermal instabilities to the thickened crust and mantle lithosphere; these instabilities are the driving force of the postcollisional evolution of the orogenic areas. According to current views, the orogens may evolve through gravitational collapse (*Rey et al., 2001*) or through mantle delaminaton (*Houseman and McKenzie, 1981*), the theories of which have been developed to explain the postcollisional magmatism and related geophysical anomalies observed in active orogens like the Cordilleras and the Himalayas. The developing crustal structures cannot, however, be studied in active orogens, but should rather be studied in older, stabilized ones with deeper erosional levels.

The Svecofennian orogen in the Fennoscandian Shield is a good example of an old (1.95-1.80 Ga), stabilized orogen with an erosional level of 15-20 km, and has been described in plate tectonic framework as long (50-100 Ma), semi-continuous collisional process similar to other Proterozoic orogenies (*Nironen, 1997; Windley, 1993*). Korja (1995) and Lahtinen et al. (2004) suggested that orogenic collapse took place after the Svecofennian orogeny. Structures related to gravitational collapse have, however, not been described from the Fennoscandian Shield, and are poorly described in Precambrian terranes elsewhere, as interpretations have been concentrating on collisional structures proving Precambrian plate tectonics.

This project will concentrate on searching for the structures related to the stabilization of the crust in the central part of the Svecofennian orogen that is occupied by the Central Finland Granitoid Complex (Fig.1), which is thought to represent a type locality of stable crust with postcollisional crustal evolution. The CFGC mainly comprises calc-alkaline I-type granitoids (1.89-1.88 Ga) with minor amounts of mafic plutonic rocks as well as remnants of deformed sedimentary and volcanic rocks, and was later intruded by a younger group of hypabyssal rocks and dikes as well as postkinematic granitoids at 1.88-1.87 Ga (*Huhma, 1986; Elliott et al., 1998; Nironen et al., 2000*). The CFGC is cross-cut by two FIRE (Finnish Reflection Experiment) deep seismic reflection lines (FIRE 1&2 and 3), whose preliminary interpretation (*Korja et al., in prep.*) indicates that the Granitoid Complex was formed on top of a stack of continental slices as a result of extension. A beautiful rift-related structure is found in the lower and middle crust. The main goal of this project is to test whether the structures found on the FIRE lines were formed by either gravitational collapse or by lithospheric delamination.

Central Finland provides an excellent possibility to combine new deep seismic reflection (FIRE) and tomographic (SVEKALAPKO) data on the structure of the lithosphere with detailed structural and petrologic studies to provide insight into the processes of crustal stabilization and formation of large igneous complexes. The Svecofennian example can be used to postulate what may happen to the modern orogenic crust and what processes will stabilize it.

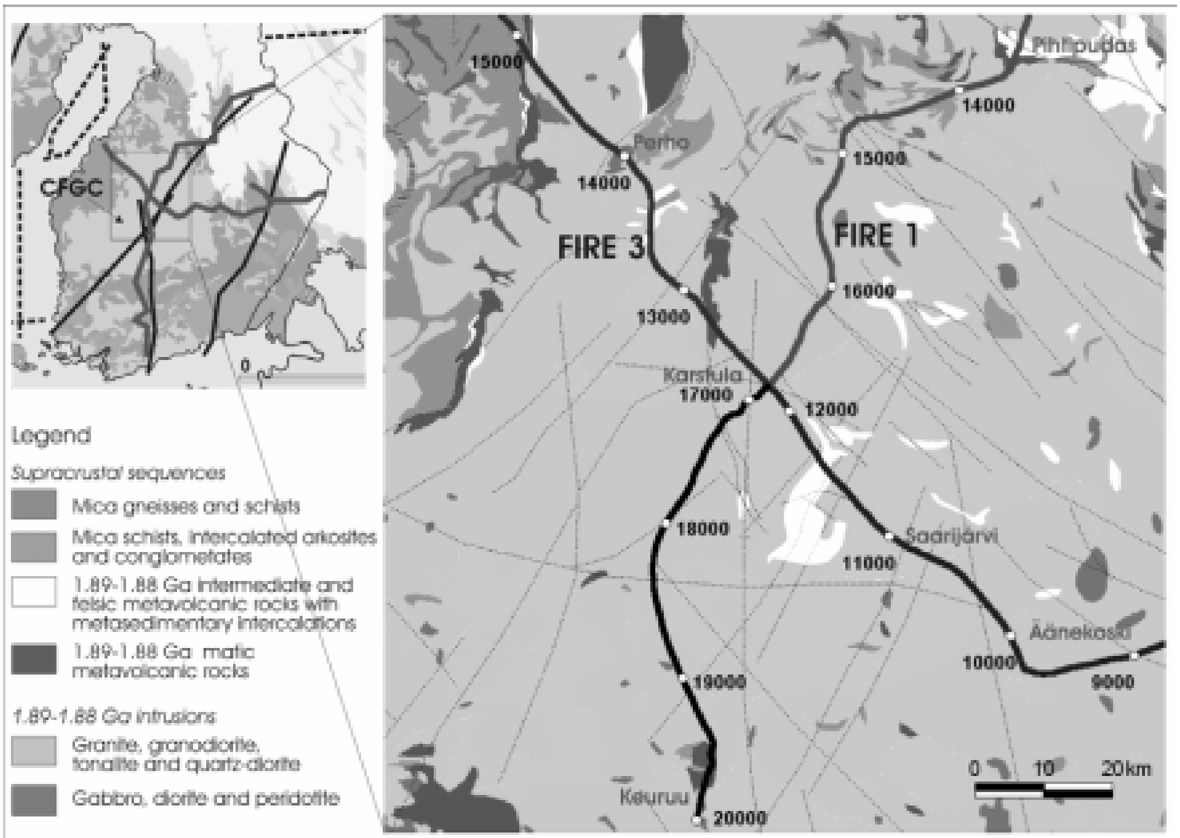

Figure1. FIRE deep seismic reflection lines 1 and 3 on the lithological map of the Central Finland Granitoid Complex. The index map after Koistinen et al. (2001), the CFGC map modified from Korsman et al. (1997).

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Structure of the upper mantle beneath central Fennoscandian Shield from seismic anisotropy studies

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The SVEKALAPKO deep seismic tomography experiment was conducted in 1998-1999 in central and southern Finland. An important part of the project is seismic anisotropy study carried out by the Institute of Geophysics of Czech Academy of Science, at present in collaboration with the Sodankylä Geophysical Observatory of Oulu University and with the Department of Geophysics of Oulu University. The study revealed several domains with different anisotropic pattern below the SVEKALAPKO seismic array. In particular, strong anisotropy and rather homogeneous orientation of anisotropic material in the upper mantle was revealed beneath Archean domain. On the contrary, the anisotropic pattern corresponding to the Proterozoic domain seems to be more heterogeneous and the anisotropy seems to be weaker in the central part.

Keywords: Lithosphere, mantle, seismic anisotropy, SVEKALAPKO

The main goal of all geophysical studies in the EUROPROBE/SVEKALAPKO multidisciplinary project (Hjelt and Daly, 1996, Hjelt and Daly, 2002) was to investigate deep structure of the lithosphere-asthenosphere boundary and structures within the lithosphere of the central part of the Fennoscandian Shield (Hjelt and Daly, 1996, Hjelt and Daly, 2002). As a part of this multidisciplinary research, the SVEKALAPKO deep seismic tomography experiment aimed at studying details of the crust and lithosphere under the central Fennoscandian Shield (Bock et al., 2001).

An important part of the SVEKALAPKO seismic array research is the seismic anisotropy studie. The main method of seismic anisotropy investigation implied in the SVEKALAPKO project is joint inversion of shear wave splitting parameters and longitudinal wave residuals aiming at retrieving the orientation of seismic anisotropic structures in the mantle lithosphere (Šilený and Plomerová (1996), Plomerová et al. (1996)). Such an approach results in retrieving a self-consistent 3-D orientation of the anisotropic structures in the upper mantle and in distinguishing between hexagonal and orthorhombic anisotropy models.

Plomerova et al. (1999, 2001, 2003) presented first results of seismic anisotropy investigations achieved by joint interpretation of P-wave residuals and S-wave splitting parameters beneath the central Finland. They demonstrated the differences in lithosphere thickness beneath the TOR and SVEKALAPKO arrays, similarly to findings of 3-D isotropic tomography, receiver functions or surface wave analysis (Bock et al., 2001). The result also indicated clear difference in anisotropic structure of Archaean and Proterozoic domains below the SVEKALAPKO area with several smaller regions in each of them.

In 2003-2004 the seismic anisotropy study continued in collaboration with the Department of Geophysics and Sodankylä Geophysical Observatory of Oulu University. The project was supported by the Academy of Finland and the Academy of Science of Czech Republic via the system of exchange visits grants. During this period the shear wave splitting data were analyzed and new methods and software of joint inversion of anisotropic shear wave splitting parameters and longitudinal wave residuals has been developed by L. Vescey and E. Kozlovskaya (Kozlovskaya et al., 2003). The method was tested with the data of the SVEKALAPKO and several other previous projects. At present, the code was applied to interpretation of the SVEKALAPKO body wave data**.** The study confirmed several domains with different anisotropic structure below the SVEKALAPKO seismic array. In particular, strong anisotropy and uniform orientation of anisotropic material in the upper mantle was revealed beneath Archean domain. On the contrary, the anisotropic pattern corresponding to the Proterozoic domain is more heterogeneous with weaker anisotropy in the central part. This may result from different orientation of mantle material within different tectonic units of this region and their collision. Alternatively, it may be an artefact resulting from weak anisotropy and/or from insufficient azimuth coverage of events in S-wave splitting data. Therefore, further analysis of shear wave splitting aiming at improving the azimuth coverage will provide more detailed information about seismic anisotropy of individual Proterozoic domains.

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Appendix

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Thermal modelling of crustal stacking and exhumation during the Palaeoproterozoic orogenic growth of the central Fennoscandian Shield

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We present results on numerical modelling of the thermal evolution of the crust in the central Fennoscandian Shield during the Palaeoproterozoic orogeny (c. 1.92-1.87 Ga). The models suggest that the crustal evolution can be successfully modelled by stacking of crustal columns which simulate the collsional and accretionary processes. Stacking is followed by warming of the lithosphere by conductive heat transfer and natural radiogenic heat production of the rocks. On the other hand, exhumation of the crust leads to cooling of the lithosphere due to removal of radiogenic heat sources from the lithosphere. Our results indicate that conductive heat transfer and the distribution of radiogenic elements in the crust are major factors controlling the thermal evolution of the crust. No large-scale magmatic heating is required. However, the transport of radiogenic elements to the upper crust by melts originating from lower and middle crust is an essential mechanism leading to final thermal stabilization of the lithosphere.

Keywords: lithosphere, thermal evolution, numerical models, Fennoscandian Shield

1. Introduction

Thermal evolution of the continental crust in the Fennoscandian Shield during the Palaeoproterozoic orogeny (c. 1.92-1.87 Ga) was modelled with one-dimensional stacking and exhumation models. The crustal evolution of the central part of the Fennoscandian Shield was characterized by break-up of the Archaean craton (c. 2.2-2.0 Ga ago), opening of an oceanic basin to the west of the craton, and subsequent closure of the oceanic basin and accretion of arcs and microcontinents to the western margin of the craton at about 1.92-1.87 Ga ago. The Archaean wedge-shaped cratonic margin was both overthrust by a 15-20 km thick stack of Paleoproterozoic arc material and underthrust by 20 km of mafic underplate. Today, these events are observed, e.g., in the reflection seismic results by the Finnish Reflection Experiment (FIRE project) and in the lower amphibolite facies overprinting at the present erosion level of the Archaean craton. Further to the west in the Proterozoic part the crustal accretion was characterized by stacking of small plates of continental material as well as related granitoid magmatism.

2. Thermal modelling

Thermal models were constructed using data on present thermal conductivity and heat production data as well as crustal structures revealed from seismic data. Crustal stacking and exhumation were simulated with 1-dimensional numerical heat conduction models in which the thermal regime was calibrated with heat production estimates based on geochemistry. In the models, the decay of radioactive elements with time as well as geological and isotope data on thermal constraints on the thermal evolution of bedrock in the area were accounted for.

3. Results and conclusions

The results support continental accretion by stacking processes. Rapid exhumation rate and partial melting events that facilitated the transport of elements responsible of radiogenic heat generation (U, Th and K) to upper crustal levels were crucial factors in the thermal stabilization of the newly accreted continental crust.

The P-T-t paths constructed with the modelings are in good agreement with the documented geological evolution of the area. The results provide further constraints on the present composition and metamorphic grade of crustal units, such as degree of eclogitization of the lowermost crust.

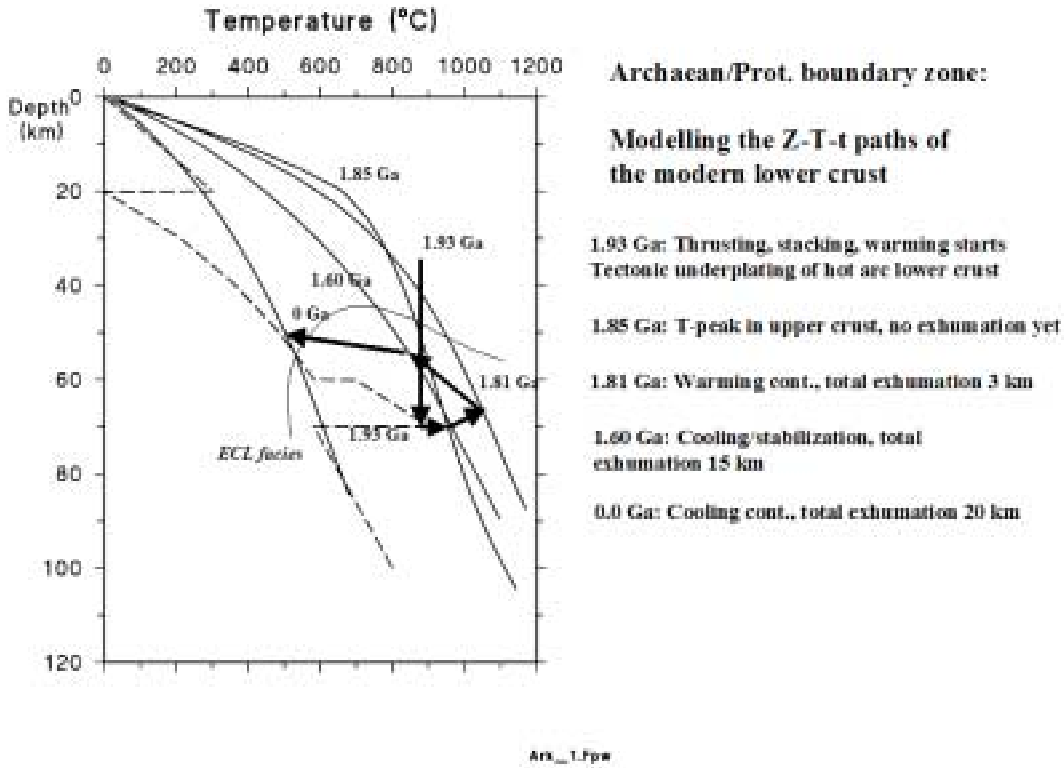

Figure 1. An example of the thermal modelling of the lithosphere in the Fennoscandian Shield. The model simulates the Palaeoproterozoic thrusting of metasediments on top of the Archaean craton and simultaneous underthrusting of arc-type warm lower crust at the Moho level. Geotherms are presented for different moments of time during the orogeny. Thick arrows indicate the depth-temperature path experienced by the (present) lower crustal rocks.

Single zircon U-Pb age results from the late Svecofennian microcline granites, southeastern Finland

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Keywords: lateorogenic, granite, U-Pb chronology, secondary ion mass spectrometry, zircon

1. Introduction

In southeastern Finland, the belt of the late Svecofennian microcline granites extends over the boundary of the Svecofennian and Karelian crustal domains. In order to draw conclusions about the source rocks of the granites and to date the collision of the arc complex of southern Finland in this area, we have carried out ion microprobe zircon analysis on five granite samples along the eastern part of this granite belt. Previous TIMS U-Pb datings are somewhat controversial, suggesting 1.80-1.82 Ga emplacement ages for these granites (*Nykänen, 1983; Vaasjoki & Mänttäri, 2000*; unpublished data by Vaasjoki & Kurhila).

2. Sample material

Our sampling sites were 1) A1721 Kaisavuori in Valkeala between the Vyborg and Ahvenisto rapakivi plutons, 2) A1751 Pulkkila in the middle of Sulkava thermal dome, 3) A1724 Pihlajaniemi at the margin of the Puruvesi batholith and 4) A1711 Rastiniemi at the center of the same Puruvesi batholith. In addition, an old sample A1307 Jaani from Mäntyharju adjacent to the Ahvenisto rapakivi granite pluton was included, mainly to clarify the somewhat heterogeneous results of a previous TIMS dating by *Vaasjoki & Mänttäri (2000)*. All the sampled granite bodies are in many ways similar to each other. They are peraluminous and contain variable amounts of garnet, features that are commonly associated with S-type granites.

The Kaisavuori granite is a relatively homogeneous and coarse-grained, garnet-rich granite. It is rich in rutile and magnetite, some of which is transformed into hematite giving the rock a reddish colour. The rock has a very weak and varying magmatic orientation, which is best seen in the sparse K-feldspar megacrysts.

The sample of Pulkkila granite is from the central part of the Sulkava thermal dome *(Korsman et al., 1984)* west of Sulkava town. The rock differs petrographically from other samples in this study. It is more plagioclase-rich, has less organized microcline structure and is devoid of monazite, which is a very abundant accessory mineral in all the other samples.

The Pihlajaniemi granite represents the marginal phase of the Puruvesi batholith. It is almost white, with irregularly dispersed garnet and abundant mica gneiss inclusions, which have been reworked to various extents. The rock also grades to pegmatite in places, and is altogether quite inhomogeneous.

The sample of Rastiniemi comes from the central phase of the Puruvesi batholith. It is porphyric, with prominent K-feldspar megacrysts and hardly any garnet. In contrast to the marginal area of the batholith, the granite of the whole central part is homogeneous.

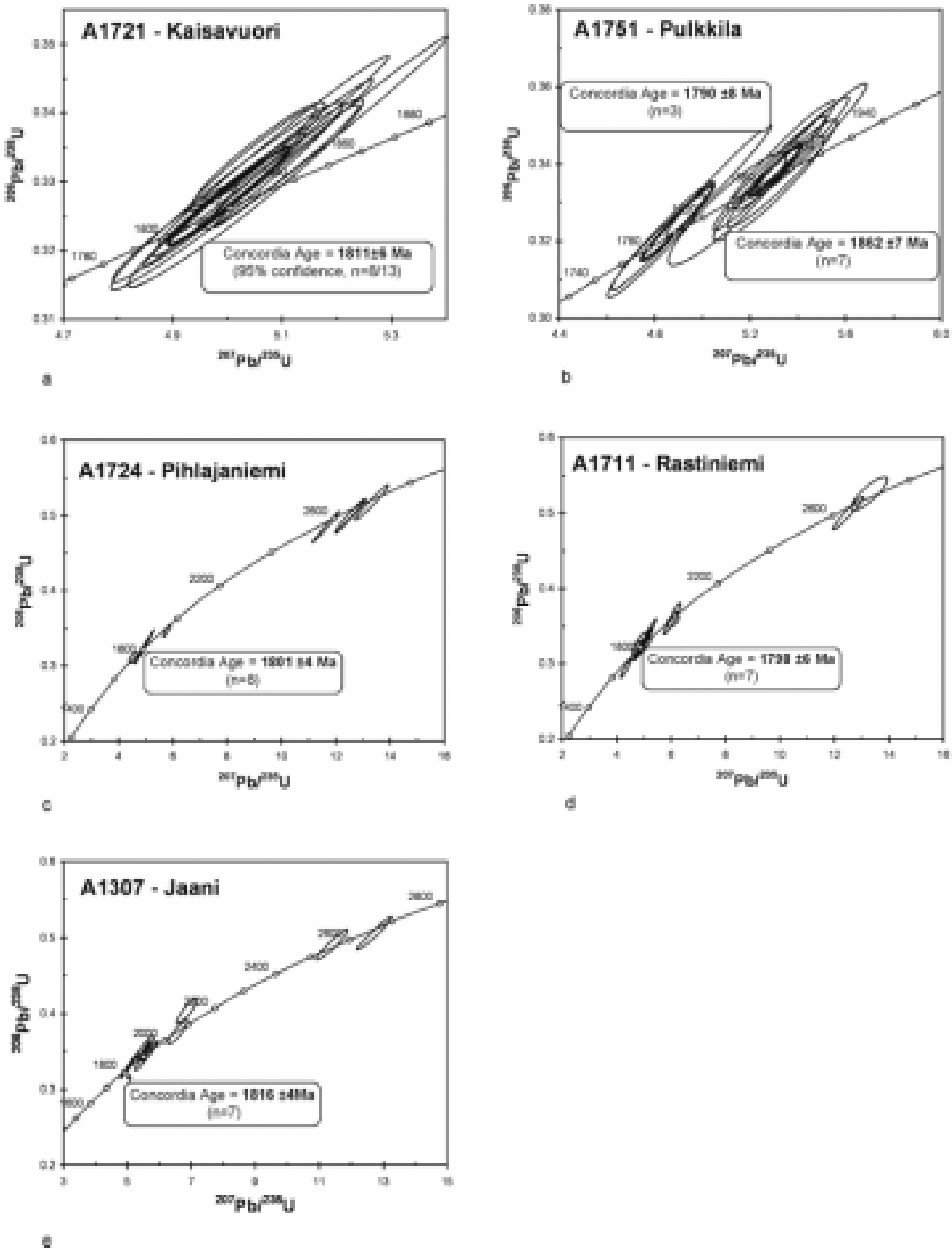

Figure 1. Concordia plots showing the ion microprobe U-Pb zircon data on the studied samples. Discordant analyses are excluded. The errors are at 2 sigma level, unless otherwise indicated. Decay constant errors are ignored.

3. Age results

The U-Pb age results are collectively presented in figure 1. In most cases, they confirm the existing conventional U-Pb ages, though some new aspects did emerge.

The Kaisavuori granite has a rather uniform zircon population with an age of 1811±6 Ma (fig. 1 a). However, if we exclude three small possibly metamorphic zircon crystals and one rim of a prismatic crystal, the age becomes 1820±10 Ma. The metamorphic zircons would then have a discordia age of 1804±9 Ma.

The zircon ages from the Pulkkila granite form two clusters (fig 1 b). The older one approaches, at 1862±7 Ma, the age of the final stages of the Svecofennian orogeny *(eg. Vaasjoki, 1996)*. As the Sulkava migmatite dome is thought to have formed by ascent of a mixture of melted and unmelted metasediments *(Korsman et al., 1984)*, this age is interpreted as the age of the source rock. The younger age, 1790±8 Ma, dates the cooling of the ascended melt. The result is slightly younger than previous ones dating the thermal event of Sulkava area *(cf. Korsman et al., 1984; Vaasjoki & Sakko, 1988)*, although the 1796±5 Ma monazite age from Härkälä metapelite *(Vaasjoki & Sakko, 1988)* north of Sulkava overlaps with our present result.

Although the two types of granites of the Puruvesi area are somewhat different in terms of petrography and geochemistry, their ages and zircon inheritance patterns are almost identical (fig. 1 c and d). The ages of magmatic zircons in both types cluster at about 1800 Ma. In addition they contain some older, both Proterozoic and Archean inherited zircon cores and whole crystals. It is therefore evident that the source rocks of the Puruvesi pluton were essentially sedimentary rocks composed of both Proterozoic and Archean detrital material.

The conventional U-Pb zircon dating of the Jaani granite in Mäntyharju by *Vaasjoki & Mänttäri (2000)* had pointed to a heterogeneous zircon source material. In their study, a discordia line defined by four fractions, consisting of zircons of obviously mixed ages, resulted in an age of approximately1880 Ma. Monazites were considerably younger at 1822 ±1 Ma. The secondary ion mass spectrometry reveals a multi-stage origin of the source material for the Jaani granite (fig 1 e). The main phase has almost the same age as the monazites, 1816±4 Ma. There is also a considerable amount of analyzed spots that plot between 1850 and 2100 Ma. Even some Archean zircon crystals were detected. The presence of Archean material is surprising, as there are no Archean country rocks anywhere near the sample site. The result argues strongly for a sedimentary origin of the source rock.

4. Conclusions

The emplacement of the lateorogenic granites of southeastern Finland occurred roughly between 1820 Ma and 1800 Ma. The extensional zone assumed to have produced the granites seems to have propagated eastwards. An exception to this pattern is the Sulkava granite, which is related to a later thermal event that ceased at about 1790 Ma. The considerable amount of inherited zircons suggests the granites were formed from sedimentary source rocks.

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Composition of the crust in the central Fennoscandian Shield: Lithological modelling of seismic velocity data

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In this study we report the first results of an investigation aiming at the lithological interpretation of the crust in the central Fennoscandian Shield using seismic wide-angle data and literature data on P-velocities of different rock types. The velocities adopted from wide-angle velocity models were compared with laboratory measurements of different rock types corrected for the crustal PT conditions in the study area. The velocity data indicate that the P-velocity does not only increase step-wise at boundaries of major crustal layers, but there is also a gradual increase of velocity within the layers. On the other hand, the laboratory measurements of velocities indicate that no single rock type is able to provide the gradual downward increasing trends. Thus, there must be gradual vertical changes in rock composition. The downward increase of velocities indicates that the composition of the crust becomes gradually more mafic when the depth increases. We have calculated vertical velocity profiles for a range of possible crustal lithological compositions.

Keywords: seismic velocity, crust

1. Introduction

Seismic velocities of crystalline rocks are controlled mainly by mineralogical composition, pressure and temperature. To a minor degree, factors such as rock texture, anisotropy, porosity and fracturing also influence the velocities. Literature data on velocities measured as functions of temperature and pressure are available for a large number of different rock types. On the other hand, high quality velocities from wide-angle data are also available on many areas, but the lithological interpretation of the crustal velocities is rarely attempted. In this study we report the first results of an investigation aiming at the lithological interpretation of the crust using seismic data in the central part of the Fennoscandian Shield. The study area is challenging because of the anomalously thick crust (52 km on average), a thick high-velocity lower crust, and abundance of granitoids in the upper crust.

2. Seismic velocity data

The study area is well covered by previous wide-angle studies, and in the present work we apply the results from SVEKA'81, SVEKA'91, FENNIA, BALTIC and POLAR profiles (*Luosto, 1997 and references therein*). P-wave velocity data were adopted from the original ray tracing models. When plotted as functions of depth, the velocities form distinct layers step-wise increases at layer boundaries as is expected for any crustal section. However, within the major crustal layers, conspicuous trends of increasing velocities are observed. In order to avoid modelling artefacts to be interpreted in this connection, we have carefully analysed the original data.

Laboratory measurements on velocities measured as functions of temperature and pressure are available in literature for different rock types. Here we applied velocities, and pressure and temperature derivatives taken from *Christensen and Mooney (1995)*. The laboratory velocities were corrected for the Finnish conditions using a xenolith-calibrated geotherm (*Kukkonen et al., 2003*) representative for the central Fennoscandian Shield.

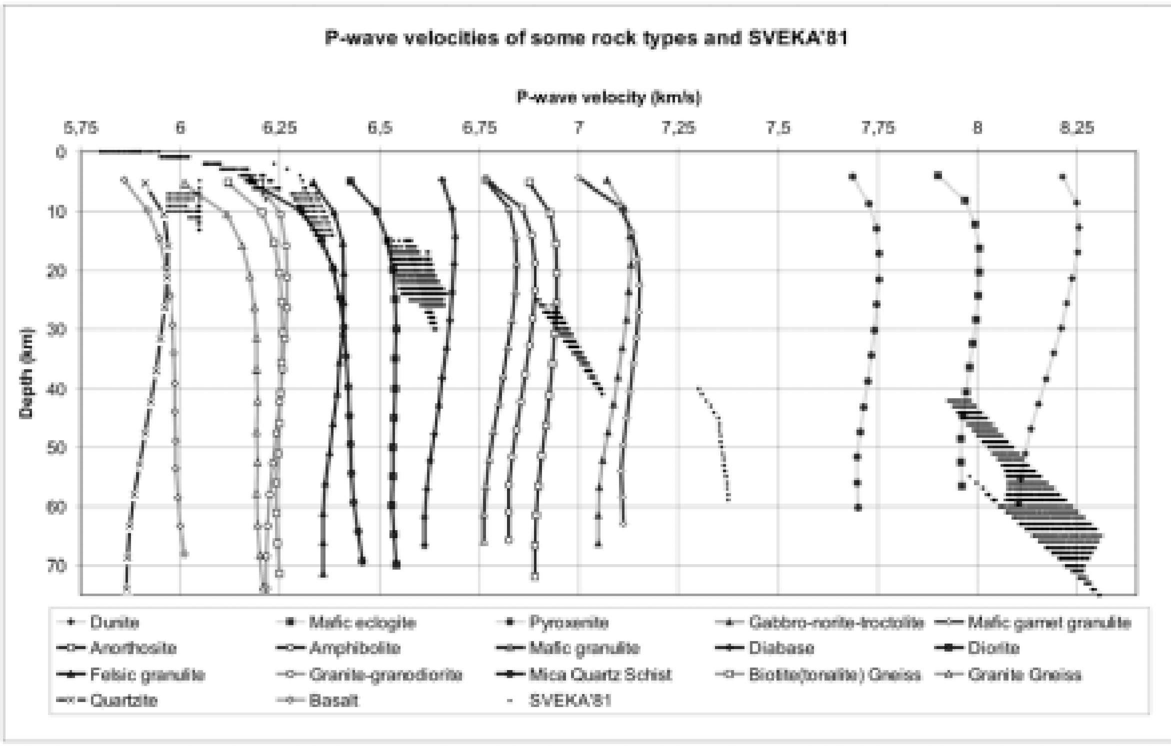

Figure 1. P-wave velocity of some rock types in PT conditions of the Finnish crust together with wide-angle model data from SVEKA'81.

3. Preliminary results

The wide-angle model velocities were compared with laboratory measurements of different rock types (an example is shown in figure 1). The results indicate that no single rock type is able to provide the downward increasing trends within different crustal layers: assuming that the velocities from wide-angle models are representative of the real *in situ* velocities, there must be gradual vertical changes in rock composition. It indicates that the composition of the crust becomes gradually more mafic within the major crustal layers, and that the compositional changes in the crust are not limited only to major layer boundaries.

The lithological interpretation of crustal velocities is a non-unique problem if only seismic velocity data is available. Nevertheless, to constrain the possible lithological types in different crustal layers, we have calculated velocities of different plausible mixtures of rock types. Here it should be taken into account that the available laboratory measurements limit the possible rock type assortment. The first results suggest that the upper crust (about 0-15 km) is reasonably well represented by a mixture of the major rock types outcropping on the surface (granitoids, felsic gneiss and a little of mafic rocks). The upper part of the middle (about 15-30 km) crust is bracketed between felsic and mafic granulite, diorite, amphibolite, diabase and anorthosite, and a mixture of them and felsic rocks is suggested for this depth layer. In the lower part of the middle crust (about 30-40 km), the rocks can be represented as a mixture of amphibolite, gabbroic rocks, and mafic garnet granulite. The lower crust (about 40-60 km) can be interpreted as a mixture of mafic garnet granulite and mafic eclogite.

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New zircon and monazite ages and Nd isotope data on the Archean complex in Koillismaa, eastern Finland

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The Archean complex in Koillismaa, eastern Finland, mainly consists of TTG-association gneisses that are at least 2.8 Ga old and younger ca. 2.7 Ga granitoids. Both lithologies contain 2.695 Ga monazite. The age and Nd isotope composition of the Archean rocks in Koillismaa area are consistent with data acquired from other parts of the Archean Karelian province.

Keywords: Koillismaa, Archean, Karelian province, U–Pb geochronology, monazite, Nd isotope

1. Introduction

The Koillismaa area forms the northernmost part of the eastern Finland Archean block that belongs to the Karelian Province of the Fennoscandian shield. The area consists of late Archean rocks that are intruded by younger units such as the early Paleoproterozoic layered gabbros (*Alapieti, 1982*) and related silicic rocks (*Lauri, 2004*) and several generations of mafic dikes (*Vuollo, 1994*). The Archean complex in Koillismaa is little studied so far and age data especially is sparse; only an imprecise U–Pb zircon date of ~2.7 Ga is available from the area (*Lauerma, 1982*). The major rock types of the complex are variably deformed TTG-association gneisses that are cut and migmatized by younger granitic-granodioritic intrusions and dikes. Minor greenstone occurrences have also been found. Aeromagnetic data and preliminary mapping indicate that the area consists of several blocks with varying erosional level. In some blocks the metamorphic grade extends up to granulite facies.

2. U–Pb geochronology

Six samples were collected for U–Pb geochronology from the Archean complex in Koillismaa (Fig. 1). Two of these were from a granulite facies orthogneiss block, and four represented younger granitoids that cut the gneisses. Several zircon fractions were analyzed from all samples (*Lauri et al., 2004*). The acquired zircon ages are rather imprecise but they indicate that the granulitic gneisses are at least 2.8 Ga old. Sample A1661, with the U–Pb zircon age of 2808±20 Ma, is the oldest sample so far analyzed in the Koillismaa area (Fig. 1). The younger granites show heterogeneous zircon populations with crystallization ages around 2.7 Ga.

Monazite was found from two granulitic gneisses and two younger granites (Fig. 1). All monazites show concordant 2.695 Ga ages (*Lauri et al., 2004*).

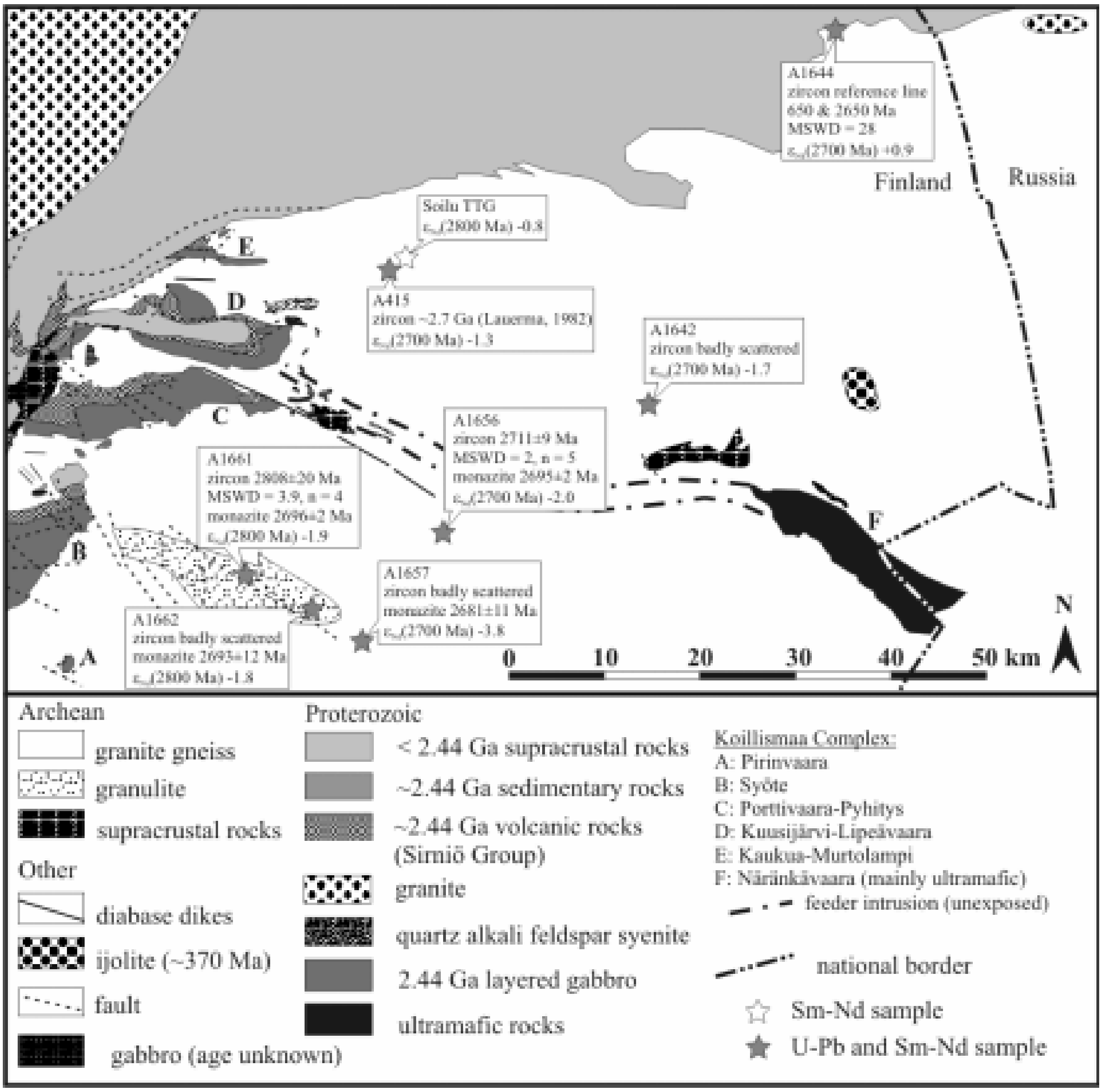

Figure 1. Bedrock map of the Koillismaa area (*modified after Räsänen et al., 2003 and Lauri et al., 2004*).

3. Nd isotope geochemistry

Eight samples from the Archean complex were analyzed for Nd isotopes (Fig. 1). The initial κ_{Nd} values (at 2800 Ma for the gneisses and at 2700 Ma for the granitoids) vary mostly between –0.8 and –3.8 and the T_{DM} model ages (*after DePaolo, 1981*) are between 2900 and 3100 Ma (*Lauri et al., 2004*). One sample (A1644) shows a markedly different κ_{Nd} value of +0.9 and T_{DM} model age of 2791 Ma (Fig. 1) that are close to the values acquired from the Archean rocks on the Russian side (*see Amelin and Neymark, 1998*).

4. Discussion and conclusions

The ages so far determined for the orthogneisses of the Archean Karelian province of the Fennoscandian shield are mainly between 2.85 and 2.68 Ga, although >3.0 Ga old rocks are also known (*e.g., Vaasjoki et al., 1999 and references therein; Hölttä et al., 2000; Hanski et al., 2001; Juopperi and Vaasjoki, 2001; Luukkonen, 2001; Evins et al., 2002; Mänttäri and Hölttä, 2002; Mutanen and Huhma, 2003*). The initial κ_{Nd} values for these rocks are between 1.6 and –4.8 and T_{DM} model ages vary between 2.75 and 3.61 Ga (*Huhma, 1986; O'Brien et al., 1993; Hölttä et al., 2000; Hanski et al., 2001; Halla, 2002*). The zircon ages and Nd isotope data acquired from the Archean complex in Koillismaa are consistent with the main Neoarchean phase of crustal growth in the Karelian Province. The 2.9 to 3.1 Ga T_{DM} ages of the Archean rocks suggest that even older rocks might be found from Koillismaa with more extensive sampling.

The monazite ages measured from the Koillismaa samples are at least some 10 Ma younger than the zircon ages of the granites and over 100 Ma younger than the zircon ages of the granulitic gneisses. The similarity of the monazite ages in both lithologies indicates a common cause for monazite growth. Although the zircon population in the granites is heterogeneous and at least partly inherited, the ~2.7 Ga zircon and monazite age is interpreted as the crystallization age for the granites. The timing of the granulite facies metamorphism in the gneisses is unknown but the monazite growth may have been induced by the heating and fluid flow associated with the intrusion of the granites at ~2.7 Ga. In some parts of the Karelian province, granulite facies metamorphism occurred at ~2.63 Ga (*e.g., Hölttä et al, 2000; Mänttäri and Hölttä, 2002*). In Koillismaa area, the Archean magmatic and metamorphic activity seems to have ceased at ~2.69 Ga. Later disturbance is negligible, as the K–Ar ages of the hornblendes and biotites are also Archean (*Kontinen et al., 1992*).

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Middle Proterozoic - Paleozoic tectono-thermal reactivation of the crust in southern Finland and northwestern Russia - paleomagnetic evidences

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Paleomagnetic investigations on shear and fault zones in southern Finland and on 1.8 Ga shoshonitic intrusions and lamprophyric dykes at Lake Ladoga area in nortwestern Russia evidence a partial or total remagnetization of the formations during the emplacement of the ca. 1.64 and 1.54 Ga rapakivi granites. In addition, the shear and fault zones in southern Finland testify repeateded reactivation during the Caledonian orogeny at ca. 420-440 Ma.

Keywords: remagnetization, paleomagnetism, shear zone, rapakivi granite, Proterozoic, Paleozoic, Finland, Russia, Fennoscandian shield

1. Introduction

Paleomagnetic studies have been carried out on shear and fault zones of the Svecofennian crust in the Helsinki capital area in southern Finland (Fig. 1). One of the purposes of the paleomagnetic study was to test the ability of remanent magnetizations to discover the relationships between different shear zones and to address timing for their evolution (*Pajunen et al., 2002*). Based on these investigations, the shear and fault zones have proven to carry multiple remanent magnetizations (*Mertanen et al., 2004a,b*). The remagnetizations were acquired mainly as a result of fluid flow within the zones of weakness, leading to the formation of new magnetic minerals. The most extensive studied shear zones, the Porkkala-Mäntsälä shear zone, has been especially vulnerable to later reactivation, evidenced by three remanent magnetization components. Two of the components can be related to the emplacement of the Subjotnian, ca. 1.64 Ga Onas and Bodom rapakivi granites. The third remanence component is suggested to originate from a much younger tectonic process during Middle Silurian at ca. 420-440 Ma, when the shear zones were reactevated as a result of the Caledonian orogenic events.

New paleomagnetic results of an ongoing investigation (*Eklund, 2003*) on 1.8 Ga post-collisional shoshonitic intrusions *(Eklund et al., 1998)* and coeval lamprophyric dykes at Lake Ladoga area (Fig. 1) also evidence Subjotnian reactivation. In addition to the primary 1.8 Ga remanence, the intrusions and dykes carry partial remagnetizations, which according to the pole positions were acquired at the time of the emplacement of the ca. 1.54 Ga Salmi rapakivi granite. Furthermore, paleomagnetic studies on undated dolerite dykes between the Salmi and Wiborg batholits imply considerable Subjotnian dyke magmatism which, thus forms an additional possible source for remagnetization. The new paleomagnetic results from the shear zones in southern Finland and from the intrusive rocks at Lake Ladoga area will be presented here.

2. Sampling and methods

In southern Finland, at the Helsinki area (Fig. 1) samples were taken from altogether 17 shear zones which were chosen according to their mineralogy, location and strike of the shear structure. Nine of the studied sites gave stable paleomagnetic results.

At Lake Ladoga area in Russia (Fig. 1), sampling was carried out on three undeformed ca. 1.8 Ga Svecofennian post-collisional shoshonitic intrusions; Vuoksa, Ojajärvi and

Elisenvaara and on five lamprophyric dykes and six doleritic dykes further northeast from the intrusions. Two of the lamprophyric dykes give consistent results, although the rest were magnetically too unstable to be used in further investigations. Three of the doleritic dykes yield a consistent, distinctive remanence direction while the remanence directions of the rest of the dolerite dykes form another directional group, and probably represent Postjotnian dyke magmatism. The results of the Postjotnian dykes are not handeled here.

Petrophysical properties; density and magnetic susceptibility, were measured for all specimens before palaeomagnetic measurements. Remanence measurements were made on 2G-Enterprises SQUID magnetometer and on Schoenstedt magnetometer. Demagnetizations were done both by alternating field (AF) up to peak field of 160 mT and thermally up to peak temperature of 680ºC. However, most of the samples from the shear zones could not be thermally demagnetized due to magneto-mineralogical alterations during heating. Magnetic carriers were identified by thermomagnetic analysis, applying stepwise heating up to 700ºC. The acquisition of isothermal remanent magnetization (IRM) was studied in four selected specimens from the shear zones that showed stable remanence directions. One to two polished thin sections were studied from each site.

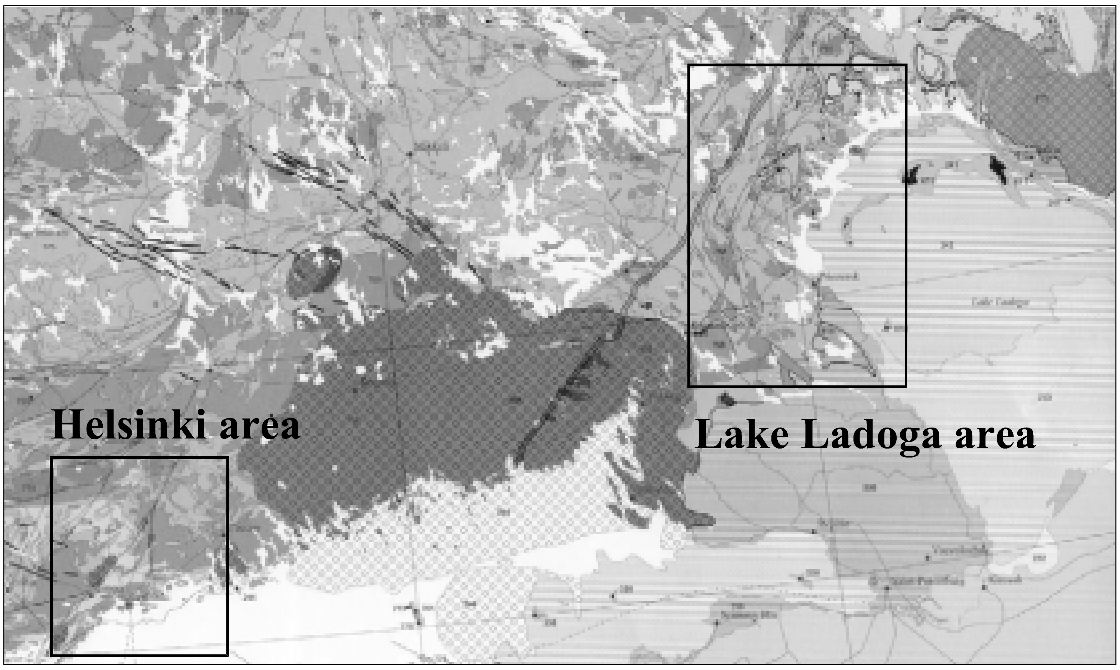

Figure 1. Geological map *(Koistinen et al., 2001)* of southeastern Finland and northwestern Russia showing the two study locations. In the Helsinki area the studies were carried out on shear and fault zones and in the Lake Ladoga area, on shoshonitic intrusions and mafic dykes.

3. Results

Shear and fault zones of the Helsinki area

Altogether five remanence components were isolated in the shear zones. The characteristic remanent magnetization, component C, was isolated only in the shear zones, but not in the host granites and gneisses. It was isolated both in low and high coercivities upon AF demagnetization. Unblocking temperatures of thermally demagnetized samples suggest that the remanence possibly resides both in titanomaghemite and titanomagnetite. Another commonly observed remanence, component SB_1, was isolated in the shear zones in unblocking temperatures below 680ºC, suggesting hematite as the main remanence carrier. The characteristic remanence component, SB_2, of the sites within the Porkkala-Mäntsälä shear zone, whether taken from the host rocks or the mylonite is extremely stable upon AF demagnetization. Thermal demagnetizations imply titanomaghemite as the main remanence carrier. A fourth remanence component B was sporadically observed in some of the mylonite and gneiss samples. Part of the shear zones and host rocks, despite of beeing in most cases too unstable to yield stable remanence directions, carry a component called A.

In order to obtain timing for the acquisition of magnetizations, the virtual geomagnetic poles (VGP's) of the remanence components were compared with the well-defined, dated Precambrian 'key poles' (Fig.2a) of the Fennoscandian Shield (see *Buchan et al., 2000*), and with the well-defined Early Ordovician to Permian segment of the Apparent Polar Wander Path (APWP) (Fig. 2b) of the Fennoscandian shield (e.g. *Torsvik et al., 1996; Smethurst, et al., 1998; Torsvik and Rehnström 2001;2003*). Based on the obtained ages, following interpretations for the origin of components are suggested.

On the basis of the closeness of pole A to the known 1.88 and 1.84 Ga key poles, the remanence component A was acquired during the Svecofennian orogenic events. Although not fully matching with the known key poles, the remanence probably represents the original Svecofennian magnetization that has preserved in part of the formations.

Poles SB_1 and SB_2 compare well with the known Subjotnian ca. 1.63 and 1.58 Ga poles, respectively. It is suggested that component SB_1 was aquired at ca. 1.63 Ga during the extensional phase at the time of the rapakivi intrusions and related dikes, when chemical or thermo chemical remanent magnetization (CRM/TCRM) was formed due to fluid migration along the faults. In that process new hematite was precipitated. Remanence component SB_2 was obtained at three studied sites, all locating within the Porkkala-Mäntsälä shear zone. The pole position suggests an age of ca. 1.58 Ga for component SB_2. The age difference of poles SB_1 and SB_2 may thus imply fluid activity and remagnetization that lasted about 50 Ma. Component SB_2 resides mainly in titanomaghemite that is implied to have been formed due to partial oxidation of titanomagnetite. Consequently, components SB_1 and SB_2, residing in different magnetic minerals were probably precipitated in two hydrothermal fluid flow events during Subjotnian at 1.63 Ga and 1.58 Ga and thus signify different compositions of fluids. On the other hand, pole SB_2 fits well to the Phanerozoic APW path, giving an age of 440 Ma. It is therefore suggested that remanence component SB_2 may also have been formed in the tectonic events related to the Caledonian orogeny.

The most common remanence component obtained in the shear zones is component C. Based on pole position on the Phanerozoic APW path, component C was acquired during Middle Silurian, at ca. 420 Ma. It is proposed that this remanence was formed due to release of tectonic stress caused by the Caledonian orogeny which resulted to the movement of fluids in the shear zones and fractures. In addition, it is implied that pervasive mild reheating event related to burial after the Caledonian orogeny is partly responsible for the remagnetization.

Pole B represents a magnetization that has been sporadically obtained over the whole Fennoscandan Shield. It does not match with any known Precambrian key poles, but it compares well with the Phanerozoic paleopoles, attesting to a Carboniferous-Permian age of ca. 300-230 Ma. It is implied that it may be related to the late sedimentary processes after the Caledonian orogeny.

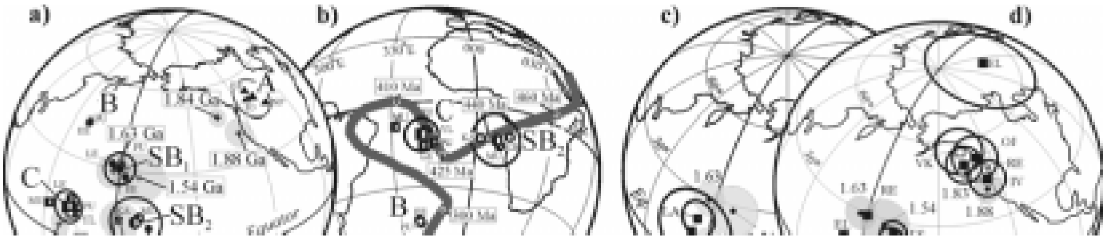

Figure 2. Mean palaeomagnetic poles for components A (triangels), B (circles), C (squares), SB_1 (quadrangles), and SB_2 (pentagons) from the shear zones. The A95 confidence circles are shown around the mean poles (no A95 values were calculated for the B pole because only two sites). Closed (open) symbols denote normal (reversed) polarity and half opened symbols mixed polarity.
a) The crosses with shaded A95 circles denote the 'key poles', the ages of which are indicated (see *Buchan et al., 2000, Pesonen et al., 2003*).
b) Fennoscandian APW path (*Torsvik, et al., 1996, Smethurst et al., 1998; Torsvik and Rehnström 2001;2003*) with relevant ages indicated. Components B, C and SB_2 are switched to opposite polarity fom their positions in Fig. (a).
c) Poles VK, OJ, EL and ES are from the Vuoksa, Ojajärvi and Elisenvaara shoshonitic intrusions, poles IV and RE from lamprophyric dykes at Lake Ladoga area.
d) Poles LA, LB an LC are from doleritic dykes at ake Ladoga area. See text.

Intrusions and dykes of the Lake Ladoga area

The quartz monzonite of the large Vuoksa intrusion (1802+/-17 Ma, U-Pb, zircon, *Konopelko and Ivanikov, 1996*) has a characteristic SE declination and intermediate negative inclination direction, while the quartz monzonites of the Ojajärvi intrusion (1800-1805 Ma, K-Ar, amphibole, *Ivanikov et al. 1996*) yields an opposite direction of NW declination and intermediate positive inclination. In both intrusions, few samples show antipodal directions. The remanence resides in titanomagnetite. The calculated pole corresponds to the known 1.88-1.84 Ga key poles (Buchan et al., 2000) of the Fennoscandian Shield (Fig. 2d). The remanence directions of the Elisenvaara intrusion (1.80+/-6 Ma, U-Pb, zircon, *D. Konopelko and M. Vaasjoki, 2002,* pers.comm.), which is exposed as two pipe-like bodies, differ from the directions of the Vuoksa and Ojajärvi intrusions. One of the studied sites (EL, Fig. 2d), comprising of a variety of shoshonitic rock types, yields SE declinations with steep negative inclinations. By applying a tectonic correction, probably due to local tilting of the site, the remanence becomes aligned with the Vuoksa direction. The other Elisenvaara body (ES, Fig. 2d) of a homogenous syenite yields SW declinations with low inclinations. The calculated pole is in agreement with the 1.54 Ga key pole of the Fennoscandian Shield (Fig. 2d). Hence, this Elisenvaara body was evidently totally remagnetized during Subjotnian. A smilar remanence direction was also observed as a partial overprint in the other Elisenvaara body (EL) as well as in some samples of the Vuoksa and Ojajärvi intrusions (Fig. 2d), implying a pervasive reactivation of the intrusions during Subjotnian.

The two magnetically stable lamprophyric dykes (IV and RE, Fig. 2d) show typical Svecofennian remanence directions (component A) that correspond to the remanence directions obtained in the Ojajärvi and Vuoksa intrusions. In these dykes the remanence shows only normal polarity. The main remanence direction is isolated below the maximum unblocking temperature of ca. 580ºC suggesting that the remanence resides in titanomagnetite. Both lamprophyric dykes carry another remanence component, S, as a partial overprint that is isolated by thermal demagnetizations (Fig. 3) in temperatures below 370ºC, suggesting titanomaghemite as the remanence carrier. This direction corresponds to the partial overprint direction obtained in the Vuoksa and Ojajärvi intrusions, and as the characteristic component in the Elisenvaara intrusion. Based on the well-defined pole position of the remanence of dyke IV (Fig. 3), the remanence was acquired during Subjotnian, at ca. 1.54 Ga.

Three of the dioritic dykes at Lake Ladoga area (Fig. 2c) carry their own characteristic remanence direction. The remanence resides in titanomagnetite. The poles compare rather well, although not fully matching with the 1.63 - 1.54 Ga key poles of the Fennoscandian Shield (Fig. 2d). Based on these ages, it is suggested that the dioritic dykes represent dyke magmatism that took place during the time between the emplacement of the Wiborg and Salmi rapakivi batholits. Emplacement of the diorite dykes form a plausible source for the partial remagnetization of the 1.8 Ga old shoshonitic intrusions and lamprophyric dykes.

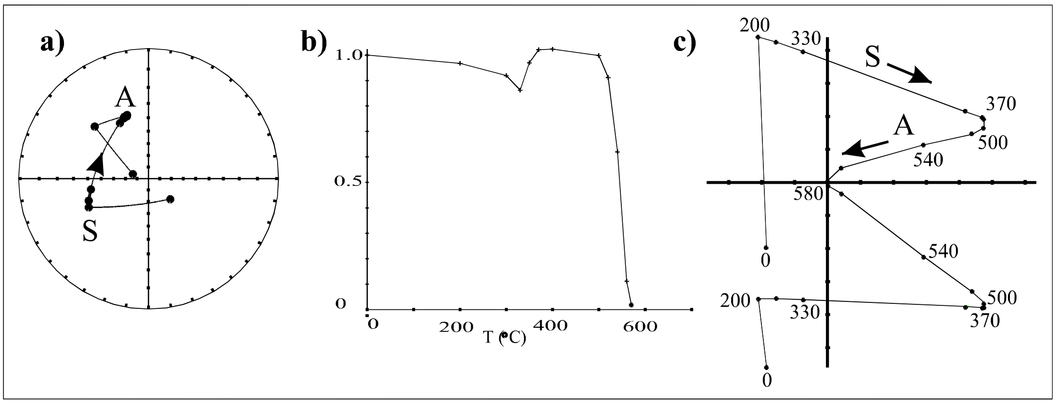

Figure 3. Thermal demagnetization behaviour of a lamprophyric dyke (IV) from Lake Ladoga area. The occurrence of a lower temperature component S and a higher temperature component A is visible in the steroplot (a), intensity decay curve (b) and Zijderweld plot (c).

4. Conclusions

Paleomagnetic studies both in southern Finland and northewestern Russia evidence the vast effect of the Subjotnian magmatism into the adjacent rocks. In both areas, a Subjotnian remanent magnetization was isolated as a partial overprint in rocks that are not in close contact to the rapakivi granites. In southern Finland the shear zones have acted as channels for fluids that probably originated in the process of emplacement of the rapakivi granites, and were able to precipitate new magnetic material. At Lake Ladoga area, the thermal effect of rapakivi granites may have extended to the intermediate areas between the rapakivi batholits, and, in addition, in this area the emplacement of the coeval Subjotnian dykes probably form another source for reactivation. Furthermore, the paleomagnetic results imply that the youngest tectonic event observed in the shear and fault zones of southern Finland is related to the Caledonian orogenic and sedimentary events.

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The Fennoscandian Land Uplift Gravity Lines: Status 2004

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The Fennoscandian Land Uplift Gravity Lines consist of four east-west profiles across the Fennoscandian postglacial rebound area, along the approximate latitudes 65º, 63º, 61º, and 56ºN. Repeated relative gravity measurements have been performed 1975–2000 (65ºN), 1966–2003 (63ºN), 1976–1983 (61ºN), and 1977–2003 (56ºN). The line 63ºN has most observations. From the measurements along it up to 1993, *Ekman and Mäkinen (1996)* deduced the ratio –0.20 µgal/mm between surface gravity change and uplift relative to the Earth's center of mass. Since that time, more gravity measurements have been taken. On the eastern part of the line 63ºN, they result in slightly smaller estimates for the rate of gravity. New estimates of uplift and model predictions are also available. The updated gravity change combined with various estimates of uplift gives ratios between –0.16 and –0.20 µgal/mm. On the western part of the line 63ºN an apparently anomalous change in gravity difference requires further study. In the future, the measurements will be performed using absolute gravity techniques.

Keywords: Postglacial rebound, gravity change, Fennoscandia, glacial isostatic adjustment

1. Introduction

The Fennoscandian Land Uplift Gravity Lines are high-precision relative gravity profiles across the Fennoscandian postglacial rebound (PGR) area (Figure 1). They run east-west, approximately along the latitudes 65º, 63º, 61º, and 56ºN. The concept and methodology were largely developed by the late A. Kiviniemi (1930–2004), who started measurements on the Finnish section of the line 63ºN 1966 *(Kiviniemi, 1974)*. Other sections *(Pettersson, 1974)* and other lines followed (Table 1). The work has been performed in Nordic cooperation (with guests from institutions around the world) and coordinated through the Nordic Geodetic Commission (http://www.nkg.fi), in particular through its Working Group for Geodynamics. *Kiviniemi (1974)* and *Mäkinen et al. (1986)* described the measurement methods. *Mäkinen et al. (1986)* gave the station descriptions, an account of the computation methods, and a detailed catalogue of results until 1986.

Table 1. Relative gravity campaigns on the Fennoscandian Land Uplift Gravity Lines 1966–2003. FI=Finnish section.

Line	Full campaigns	Partial campaigns
65ºN	1975, 1980, 1981	1999, 2000 (FI)
63ºN	Every 5 years 1966–2003	Every 1–3 years 1966–2003 (FI)
61ºN	1976, 1983	
56ºN	1977, 1984, 2003	

When the work started, the general features of the vertical motion associated with the Fennoscandian PGR were already rather well known. The research agenda was to determine $\dot{g}$, the rate of change in surface gravity, and compare it with $\dot{h}$, the rate of crustal uplift. The ratio $\dot{g}/\dot{h}$ would then allow conclusions on the physics of the uplift *(Kiviniemi, 1974)*.

This paper reviews the status of the work in 2004. An expanded version appears elsewhere *(Mäkinen et al., submitted)*.

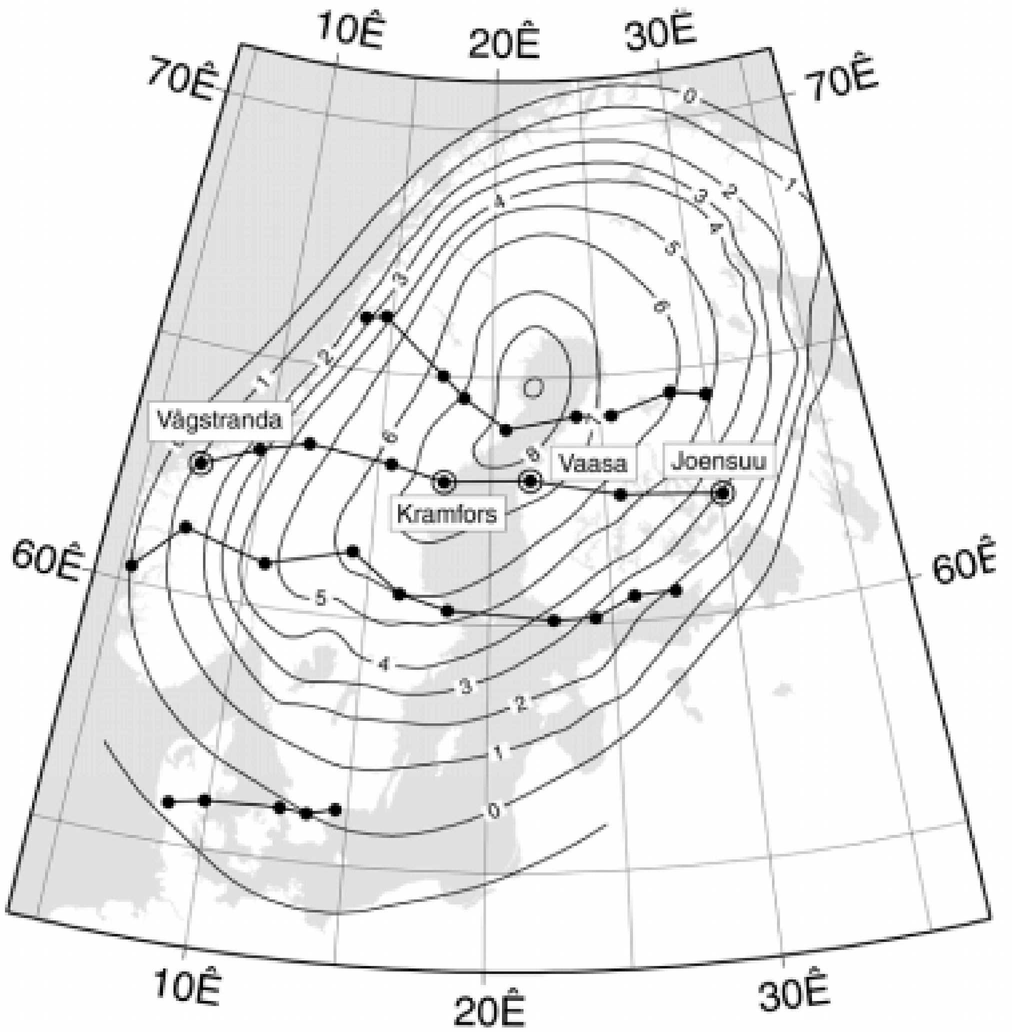

Figure 1. The Fennoscandian Land Uplift Gravity Lines. The named stations on the line 63ºN are used in this paper. The isolines show postglacial rebound rates (mm/yr) relative to mean sea level according to *Ekman* (1996).

2. Uplift and gravity change

Two simple geometric images of the uplift process have often been used to construct bounds for $\dot{g}/\dot{h}$ *(Honkasalo and Kukkamäki, 1964)*. Assume first that the uplift is due to decompression with no additional mass. Then at a point above the surface of the Earth and fixed relative to its centre of mass gravity is not changing at all. At a point moving with the surface of the Earth, gravity change can be obtained with the free air gradient. This leads to the "free air model" $\dot{g}/\dot{h} = -0.31$ µgal/mm. (1 µgal = 10^{-8}m/s^2).

Suppose, second, there is no decompression, but an inflow of mass in the upper mantle is pushing the crust upwards. Then at a point slightly above the surface of the Earth and fixed relative to its centre of mass, gravity increase can be approximated by the attraction of a Bouguer sheet with the density of the upper mantle (3300 kg/m^3). At a point moving with the surface, this leads to $\dot g / \dot h = -0.17$ µgal/mm, the "Bouguer model".

This latter figure is close to the relationship (–1 µgal)/(6.5 mm) = –0.15 µgal/mm, obtained numerically by *Wahr et al. (1995)*. They used several models of the glacial isostatic adjustment (GIA), all with Maxwell rheology of the mantle but different viscosity profiles. See also *Fang and Hager (2001)*.

In the discussion above, $\dot h$ refers to the Earth's centre of mass ("absolute uplift"). GPS observations give absolute uplift rates. Uplift rates from tide gauges refer to the mean sea level ("apparent uplift"), and rates from repeated precise levelling to the geoid. We thus have

$$\dot h = \dot H + \dot N \qquad (1a)$$

$$\dot H = \dot H_a + \dot H_e \qquad (1b)$$

Here $\dot H_a$ is apparent uplift, $\dot H$ is uplift relative to the geoid, $\dot N$ is the uplift of the geoid, and $\dot H_e$ (neglecting changes in sea surface topography) the eustatic rise in mean sea level.

In this paper we are concerned with uplift differences between stations. From them, $\dot H_e$ is eliminated, i.e. $\dot H_a$ differences and $\dot H$ differences are identical. However, $\dot H$ differences need to be corrected for $\dot N$ to obtain $\dot h$. *Ekman and Mäkinen (1996)* estimated (using their observed $\dot g / \dot H$) that $\dot N$ differences are roughly 6% of $\dot H$ differences. The GIA model by *Milne et al. (2001)* gives approximately the same percentage.

To avoid a clogged notation, we use $\dot g, \dot h, \dot H, \dot N$ to refer both to values at sites and to differences between sites. It will be clear from the context, which one is intended.

3. Previous estimates

Ekman and Mäkinen (1996) divided the line 63°N into two parts (Figure 1): the western part Vågstranda–Kramfors (W), and the eastern part Vaasa–Joensuu (E). Writing symbolically with conventional one-sigma standard errors

$$(\dot g/\dot h)_{obs} = \frac{-1.52 \pm 0.20\ \mu\text{gal/yr}}{6.9 \pm 0.5\ \text{mm/yr}} \quad \text{(W)} \qquad (2a)$$

$$(\dot g/\dot h)_{obs} = \frac{-1.00 \pm 0.14\ \mu\text{gal/yr}}{-4.7 \pm 0.5\ \text{mm/yr}} \quad \text{(E)} \qquad (2b)$$

With the relevant Student's t-distributions the ratios correspond to the 95% confidence intervals

$$(\dot g/\dot H)_{obs} = -0.220 \pm 0.086\ \mu\text{gal/mm} \quad \text{(W)} \quad (3a)$$

$$(\dot g/\dot H)_{obs} = -0.213 \pm 0.080\ \mu\text{gal/mm} \quad \text{(E)} \quad (3b)$$

After an iterative calculation of $\dot N$ to obtain $\dot h = \dot H + \dot N$, *Ekman and Mäkinen (1996)* found the corresponding intervals for $\dot g/\dot h$

$$(\dot{g}/\dot{h})_{obs} = -0.208 \pm 0.086 \ \mu\text{gal/mm} \ \ (W) \ \ (4a)$$

$$(\dot{g}/\dot{h})_{obs} = -0.200 \pm 0.080 \ \mu\text{gal/mm} \ \ (E) \ \ (4b)$$

The combined 95% confidence interval is

$$(\dot{g}/\dot{h})_{obs} = -0.204 \pm 0.058 \ \mu\text{gal/mm} \qquad (5)$$

from which they concluded that the free air model is ruled out by the observations.

The results (1975–2000) on the Finnish section of the line 65ºN were analyzed by *Ruotsalainen (2002)*. The estimated $\dot{g}/\dot{h}$ agrees with (5) but due to the paucity of observations its uncertainty is still rather large,

In the following we summarize new results of the line 63ºN since 1993 and check to which extent they modify the numbers in equations (2)…(5).

4. New gravity results

Since 1993, there have been two campaigns on the western part of the line 63ºN, in 1998 and 2003. Their results do not agree with the earlier ones (Figure 2, left). Weighted regression now gives

$$\dot{g} = -1.07 \pm 0.24 \ \mu\text{gal/yr} \qquad (W) \qquad (6)$$

where the error estimate is the conventional one-sigma. This is a large change from the previous estimate –1.52 µgal/yr (numerator of equation 2a).

There is no comparable anomaly on the eastern part of the line, with five new campaigns (Figure 2, right). Weighted regression gives

$$\dot{g} = +0.91 \pm 0.09 \ \mu\text{gal/yr} \ (E) \qquad (7)$$

again with the conventional one-sigma error estimate. This $\dot{g}$ is slightly smaller than the previous estimate +1.00 µgal/yr (numerator of equation 2b).

We are currently screening the results in more detail to understand the ostensible jump in the time series on the western part.

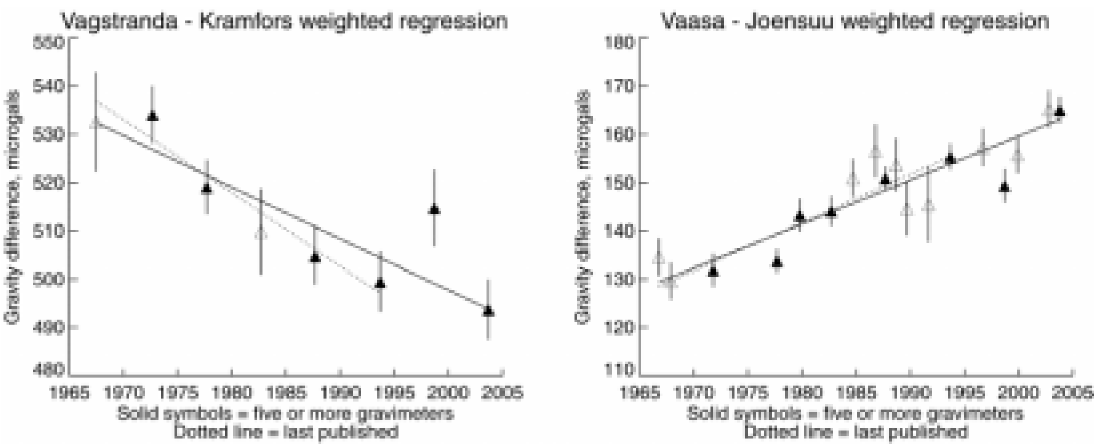

Figure 2. Results of the relative gravity measurements on the western part (left) and on the eastern part (right) of the line 63ºN. Each point is the unweighted mean of all gravimeters in the particular year. Regression lines are calculated with the number of gravimeters as weights. The solid line uses all data, the dotted line uses data until 1993.

5. New uplift results and ratios

On the eastern part of the line, the value of $\dot{H}$ in the denominator of equation (2b), i.e.,

$$\dot{H} = -4.7 \pm 0.5 \text{ mm/yr} \qquad (8)$$

results from two Finnish precise levellings, the same that underlie Figure 1. Now $\dot{H}$ values from three levellings are available *(Mäkinen and Saaranen, 1998)*. Furthermore, the land uplift stations at Joensuu and Vaasa are within 0.1 km and 2 km, respectively, of continuous GPS stations in the FinnRef® network. For these stations, $\dot{h}$ values are available both from FinnRef® *(Mäkinen et al., 2003)* and BIFROST *(Johansson et al., 2002)* processing of five years of GPS observations. We use the winter-edited results (*op. cit.*)

In addition, we employ predicted $\dot{H}_a$ values from the GIA model by *Lambeck et al. (1998)*. The main observational constraints for this GIA model are relative sea level histories, although it has been fine-tuned by a fit to the tide gauge rates computed by *Ekman (1996)*. The GIA model by *Milne et al. (2001)* uses the same ice history, but the rheology is fitted to 3-D velocities from the BIFROST project *(Johansson et al., 2002)*.

The uplift results and corresponding ratios $\dot{g}/\dot{h}$ are collected in Table 2. The new estimate of $\dot{g}$ and most of the new uplift estimates bring the ratio closer to the Bouguer model. On the western part of the line the gravity measurements and the resulting $\dot{g}$ value are still under scrutiny. Therefore we have not computed new $\dot{g}/\dot{h}$ ratios.

Table 2. The ratio $\dot{g}/\dot{h}$ on the eastern part of the line 63ºN, computed using different values of $\dot{g}$ and $\dot{h}$ or $\dot{H}$. The old $\dot{g}$ is from equation (2b) and the new $\dot{g}$ from equation (5). The sources of $\dot{h}$ or $\dot{H}$ are described in the text. The $\dot{H}$ values are marked by an asterisk. They are multiplied by 1.06 to obtain $\dot{h}$. The previous estimate of $\dot{g}/\dot{h}$ is shown in boldface.

Source of $\dot{h}$ or $\dot{H}$	$\dot{h}$ or $\dot{H}$* mm/yr	$\dot{g}/\dot{h}$ old $\dot{g}$ µgal/mm	$\dot{g}/\dot{h}$ new $\dot{g}$ µgal/mm
Two levellings 1892–1955	–4.7*	**–0.21**	–0.18
Three levellings 1892–2004	–5.5*	–0.17	–0.16
Continuous GPS, FinnRef®	–5.7	–0.18	–0.16
Continuous GPS, BIFROST	–4.8	–0.20	–0.18
Lambeck et al. (1998) GIA model	–4.2*	–0.22	–0.20
Milne et al. (2001) GIA model	–5.3	–0.19	–0.17

6. Summary and discussion

We have used new gravity observations and new uplift results on the eastern part of the 63ºN line to compute new estimates of the ratio between gravity change and vertical motion connected with postglacial rebound. On this part, the previous estimate is –0.21 µgal/mm. The new estimates are –0.16…–0.20 µgal/mm. They are thus closer to the ratio for the

Bouguer model (–0.17 µgal/mm). On the western part of the line an apparently anomalous gravity change must be further studied before any conclusions can be drawn.

For this kind of study, relative gravity measurements are less accurate and more laborious than absolute measurements. Therefore the time series will be continued with absolute gravimeters. The end points of the 63ºN line, and the Danish section of the 56ºN line have already been equipped with absolute stations. For the other lines, we are studying the possibility of using portable absolute meters directly on the outdoor sites.

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Oxygen isotope and trace element zoning in garnets from granulite facies psammopelitic migmatite, SW Finland

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Oxygen isotopes (^{16}O and ^{18}O) and trace elements in profiles over garnet were analysed using a Cameca 1270 ion probe at the NORDSIM facility in Stockholm, Sweden. The garnets are from the mesosome, melanosome and leucosome of a psammopelitic granulite facies sample. Mn shows patterns typical for high-grade garnets which have been retrogressed at rims in melanosome and mesosome whereas the leucosome garnet has a smooth pattern with some lower and some higher concentrations ca 2-3 mm inside the rim. Decreased Ti, Y and REE at rims of melanosome and mesosome garnets can be coupled with retrogression of garnet and growth of biotite and accessory phases. ι ^{18}O (SMOW) show variations of up to 7 per mil (2ω ~0.5 to 1.4 ‰) between different regions of the garnets. The patterns are different from what would be expected during garnet growth in a closed system during increasing temperature.

Keywords: oxygen isotopes, trace elements, garnet, psammopelitic migmatite

1. Introduction

Both major and trace element zoning patterns in garnets may be excellent tools in solving the metamorphic history of a sample; e.g. Mn can be used to confirm retrogression at rims of garnet (*e.g. Spear 1993*) and Y and REE zoning can be coupled with accessory phases (*Pyle & Spear 1999; Rubatto 2002; Nyström & Kriegsman 2003; Whitehouse & Platt 2003*). Garnet oxygen isotope zoning patterns can potentially be used as well, but are dependent on bulk rock composition, on the presence or absence of fluids and on the reaction and PT history (*Kohn 1993*). We analysed Mn, oxygen isotopes and trace elements on garnets situated in the mesosome, melanosome and leucosome of the same psammopelitic, stromatic migmatite sample from the Turku granulite facies area in SW Finland in order to study the behaviour of O-isotopes during partial melting and back reaction. The garnets have grown during near isobaric heating from ca. 700-800°C at 6-7 kbar (*Johannes et al., 2003*).

2. Mineralogy and textures

Thin sections of the same sample (but containing different garnet grains than the ones analysed by SIMS) show that the mesosome and melanosome commonly contain cordierite with inclusions of sillimanite whereas sillimanite is not found in garnets. Instead, the garnet commonly encloses biotite and locally cordierite, quartz, plagioclase (uncommon), opaques, monazite and zircon. The textures are interpreted as growth of cordierite followed by garnet production by Crd + Bt + Qtz (∂ Pl ∂ Kfs) = Grt + L and Bt + Qtz + Pl = Grt + L (or Kfs + H_2O). In one case inclusions of cordierite and cordierite rimming garnet in melanosome showed the same extinction angle indicating that locally cordierite and garnet have grown together. Leucosome garnets contain mainly inclusions of lobate quartz, but also some biotite can be found in the innermost regions. It is possible that part of the leucosome garnet has crystallized from melt.

Garnet in especially meso- and melanosome may be rimmed by cordierite alone but more commonly biotite, quartz and plagioclase are also present. In places biotite +

plagioclase ∂ quartz rim garnet without cordierite. These textures suggest a reversal of above reactions on the retrograde path. Small sillimanite needles (∂ biotite and quartz) at some garnet-cordierite interfaces may have formed via: Crd + Grt + L = Sil + Bt + Qtz (+ Pl). Leucosome garnet is commonly surrounded only by quartz and feldspars but in places also by biotite + quartz ∂ feldspars, indicating local retrogression.

3. Analytical methods

Slabs of garnet were cut and mounted in epoxy. *In situ* analyses of oxygen isotopes and trace elements were performed in spots along the slabs using a Cameca IMS1270 large geometry ion probe at the NORDSIM facility in Stockholm, Sweden. The oxygen isotopes (^{16}O and ^{18}O) were analysed using a Cs^{+} primary beam with a spatial resolution of 15 µm, electron beam charge-compensation, and double Faraday cup collection. External reproducibility of the UWG-2 garnet (*Valley et al., 1995*) varied between 0.5 and 1 per mil (2ω) for the different analytical sessions.

Mn and REE, Y, Sr, Th and Ti were analysed in the same spots as the oxygen isotopes using a 30 σm O_2^- primary beam and a hybrid mono/multicollection electron multiplier detection protocol (*Whitehouse, 2004*). We used NIST 610 (*Pearce et al. 1996*) as an internal standard. A value of 36.82 wt % SiO_2 in the garnets was used for calculations (average garnet SiO_2 content in the same sample). Detection limits for most elements are <10 ppb. Analytical errors range from ca. ±50% RSD for the lowest concentration elements such as La (ca. 10 ppb) to <2% RSD for higher concentration elements such as Y and Sr (ca. 100 ppm).

4. Analytical results

Some chemical patterns over mesosome, melanosome and leucosome garnets are presented in Fig. 1. Garnet core regions have relatively flat Mn patterns whereas the contents increase at rims in melanosome and mesosome garnets. In the leucosome garnet shows a low Mn content ca 2 mm from one rim, whereas this rim has similar contents as the core. At ca 1.5 mm from the other rim Mn is slightly increased but drops again towards the rim. Core regions of leucosome and melanosome garnet have higher Mn concentrations (ca 10000-12000 ppm) than mesosome garnet (ca 7000 – 8000 ppm). The patterns of Sr more or less follow manganese and correlation coefficients between Mn and Sr within the garnets are 0.58-0.67. Mesosome and melanosome garnets show a decrease in REE and Ti contents in rim regions where Mn and Sr increase. In core regions of mesosome garnet HREE first decrease, then increase before the decrease towards rims. In the melanosome garnet HREE increases away from core before decreasing at rims. In the leucosome garnet Ti rather increases at rims whereas REE are fairly constant at one rim and show lower values at the other. Average HREE contents are highest in the leucosome garnet. It also shows an increase in HREE away from core regions.

The location of garnet cores was interpreted on the basis of inclusion patterns in mesosome and melanosome garnets and based on geometry in leucosome garnet. These spots coincide with increased ι^{18}O (SMOW): mesosome 12.09 ‰ ∂ 0.92 (2ω), melanosome 11.47 ‰ ∂ 1.03 (2ω) and leuco (average of two spots) 12.79 ‰ (∂ 0.54 & 0.65 (2ω)) which start to decreases away from core regions. The mesosome garnet shows humps with increased ι^{18}O ca 2-3 mm from both rims before reaching a ca 7 per mil ‰ level at rims. In the melanosome ι^{18}O first drops by ca 1 per mil on one side, then is constant until it increases close to the rim where it reaches the core level. Towards the other rim the values drop by ca 2 per mil. The variations are small compared to errors and the oxygen isotope pattern may be interpreted as flat in the melanosome garnet. In the leucosome garnet the ι^{18}O first decreases towards rims but 2-4 mm from the rims there are humps with higher values which then decrease again reaching ca 6.4 ‰ close to one rim. At the outermost rims the value jumps to over 13 ‰. At the other rim ι^{18}O reaches ca 9 ‰. In the mesosome garnet one spot was negative and had extreme errors and is excluded from the results. Also in the leucosome garnet one spot was negative but with an error similar as in other analyses. It deviates so much from other results that it is excluded. Negative ι^{18}O may be produced in minerals by interaction with meteoric water (*Elsenheimer & Valley, 1993*) and possibly the analysis hit a crack.

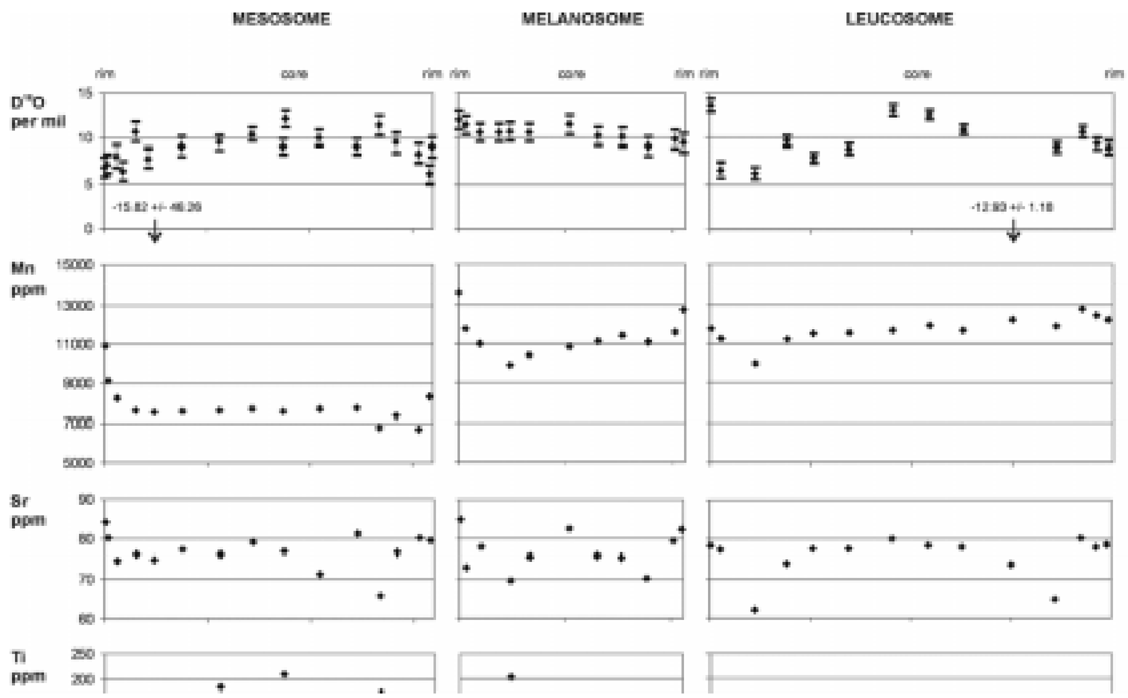

Figure 1. Chemical zoning patterns in garnet from mesosome (left column), melanosome (centre) and leucosome (right column) of a psammopelitic stromatic migmatite sample in Turku, SW Finland.

5. Discussion

Flat Mn trends in core regions of garnets is consistent with equilibration of Mn at high metamorphic grade whereas the increasing Mn at rims in melanosome and mesosome garnet is in accordance with the retrogressed appearance (*Spear 1993*). Decreased REE in the same areas where Mn increases at rims of melanosome and mesosome garnet can be interpreted by the growth of accessory phases during restite -melt back reaction between garnet and melt which is in accordance with results by *Frasier et al., 1997; Degeling et al., 2001, Pyle & Spear, 1999 and Nyström and Kriegsman, 2003*. The decrease of Ti at rims of melanosome and mesosome garnets may indicate transfer of Ti to biotite growing on the retrograde path whereas the increased (or unmodified) Ti at rims of in leucosome garnet is in accordance with a lack of biotite growth and with a partition coefficient larger for garnet than felsic melt (compilation in *Rollinson 1993*) which suggests that rims of leucosome garnet may have crystallized from melt. It is surprising that Sr follows the pattern of Mn in all analysed garnets as feldspars are common products of restite – melt back reaction and garnet has a low partition coefficient for Sr (<1) in felsic melt compared to feldspars (*compilation in Rollinson, 1993*). *Bea et al. (1994)* noticed that compared to leucosome veins K-feldspar has low Sr contents whereas plagioclase has higher which may indicate lack of retrograde plagioclase production at rims of the analysed garnets.

Generally ι^{18}O patterns start with a decrease away from interpreted core regions of garnets, opposite to what would be expected during prograde garnet growth by dehydration reactions in a closed system (*Kohn et al., 1993*). Decreasing trends in a closed system can occur in garnets growing during cooling (*Kohn & Valley, 1993*). However, the changes in our samples appear too large for cooling in a closed system and textures indicate early garnet growth during increasing temperatures. The $T_{c,\ Dodson}$ closure temperature for garnet is ca 800 ºC (*Kohn & Valley, 1998*), very close to peak metamorphic temperatures (*Väisänen & Hölttä, 1999*). The diffusion rate for oxygen in garnet is however slow (*Kohn & Valley, 1998*) and growth zoning is not likely to have been largely altered between peak and the closure temperature if it was exceeded. Oxygen isotope ratios measured by SIMS may be affected by an instrumental mass fractionation which partly depends on the mineral composition. This matrix effect can be minimized by using a standard with a composition close to the analysed sample (*Vielzeuf et al., 2003*). Since we used a garnet standard with a composition fairly close to the analysed garnets and variations in ι^{18}O are not believed to be caused by the matrix effect. Large variations with decreasing ι^{18}O may be interpreted in terms of interaction with a low ι^{18}O fluid during garnet growth and / or recrystallization (*Chamberlain & Conrad, 1993; Elsenheimer & Valley, 1993; Hoffbauer et al., 1994; Crowe et al., 2001*) and according to *Farquhar et al. (1996)* open system interactions may occur at any stage of metamorphism. *Farquhar et al. (1996)* studied a ca 900ºC granulite facies terrain and concluded that O-isotopes were changed in most minerals during high temperature open system processes, not however in garnet.

External fluids would not generally be expected in granulite facies areas as dehydrating areas below would be expected already to be dyhydrated due to higher temperature. We prefer to consider the possibilities that 1) local original variations in source composition may be large and that dehydration melting of various layers in the psammopelitic rocks has produced fluids with variable O-isotopic composition which have circulated on local scale and enhanced zoning in garnets; 2) at high metamorphic grades the presence of melt changes the distribution of oxygen isotopes between garnet and the melt compared to lower grades where dehydration has produced mainly vapour and 3) the composition of existing fluid (more CO_2 rich) may affect the O-isotope distribution. It cannot be excluded that restite – melt back reaction (with dissolved fluid released from last crystallizing melt) may have some effect on O-isotope distribution and also it must be considered that the migmatites occur together with S-type granites all of which are not *in situ* and which could have affected the oxygen isotopes by releasing fluids during crystallization.

Jumps in the trends of ι^{18}O suggest that the garnets may consist of aggregates and have not uniformly grown from one centre. As the Mn pattern within garnets is generally flat indicating equilibration at high grade these jumps may also indicate the slow diffusion of oxygen isotopes suggesting that decreasing $\iota^{18}O$ from core is a true prograde feature. Local compositions deviating from the general trends may also be due to small inclusions.

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Exploration of the Deep Subterranean Biosphere

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The exploration of deep subteranean microbial life began about 20 years ago and it has revealed an earlier unknown, vast deep biosphere [1]. Rough estimates argue that about half of planet Earth's living biomass is out of sight, dwelling below the ground surface and under the sea floor [7]. The numbers for sub-continental life has a high uncertainty and need more exploration to become accurate. The deepest subsurface location at which life can be found correlates with the maximum temperature for life (about 120 °C) and varies from a couple of meters to at least 10 km, depending on the place explored [2]. Oxygen is virtually absent just below the groundwater table, and the electrochemical potential (E_h) rapidly drops to negative values with depth. To be able to reach the intra-terrestrials we must make a hole (or a tunnel). The preferred way of scientific drilling is to use core drilling. Contamination is a threat to all microbiological sampling and it is particularly complicated to control contamination when investigating the microbiology of deep subterranean environments. It must be clarified that the organisms isolated or otherwise detected were present in the sample before drilling. Therefore, the possible contamination of sediment samples with organisms from the drill water is usually controlled by addition of dissolved and particulate tracers [e.g. 5, 6]. Once the cores have been retrieved they can be analysed for mineralogy and microbiology [3]. The hole in the ground can later be explored in several different ways. Typically, you pack off the water conducting fracture or aquifer that you are interested in. One simple way is to pump water though tubings, but with 1000 m boreholes it takes time and much happens on the way, including a pressure drop and changes in temperature. A second option is to lower samplers and open them at sampling depth. Then, the pressure will be kept at in situ level. This is the preferred option for microbiology. Finally you may need consider to bring an advanced field laboratory to the site, as samples for microbiology deteriorate quickly.

There are two main routs to explore the biogeochemistry of the subterranean biosphere. One is to sample and analyse microbial diversity, activity and distribution of active microbial life [3]. The other main rout is to analyse traces of this biosphere. We can analyse various biosignatures such as anomalies in stable the isotope composition or presence of organic molecules typical for life. A very obvious proof for subterranean life is the detection of fossils [4], provided it can be argued that the organisms of origin for the fossil were active in the explored subterranean environment.

Numbers and diversity of intra-terrestrial microbes are high, but some microbial groups, such as strict aerobes, pathogens and photosynthetic organisms are lacking for obvious reasons. The metabolic activity of deep dwelling microbes is often found to be very slow. The availability of energy and nutrients is more often than not diffusion limited in deep environments. Therefore, many intra-terrestrials are driving in one of the slowest metabolic lanes known. Who needs to rush when the circumstances during the next millennium will be almost identical to the one that just passed by? Patience and slowness are characteristics that will win over abilities to grow fast and act quickly. Such endurance qualities likely dominated among the first organisms on our planet. Those might have originated deep underground in a non-competitive environment.

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Isotopic constraints for the origin of the mafic lower crust at the Karelian craton margin

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Lower crustal xenoliths recovered from the c.600 Ma old Kaavi-Kuopio kimberlites, eastern Finland, provide us direct access to the chemical composition, physical state and age of the craton margin lower crust. We provide new ion probe U-Pb zircon dates, and Nd, Sr and Pb isotope analyses for mafic granulite xenoliths, which imply that the Karelian craton lower crust is hybrid in composition. It still contains remnants of Early Archean 3.5 Ga old granulites, which became mixed with younger underplating magmas during the Svecofennian–Archean collision, and later recrystallised by transient post-orogenic heating. Wider implications for the origin of Early Archean tectosphere will also be discussed.

***Keywords**:* lower crust, xenoliths, kimberlite, zircon, Nd-Sr, Archean, Proterozoic, Karelia, craton

1. Introduction

The lower crust of Archean cratons remains a poorly characterised realm. Although exposed Archean granulites are relatively common, it is not always evident whether they represent lower continental crust or granulites formed at transient high pressure-temperature conditions during continent–continent collisions *(Rudnick, 1992)*. Therefore, xenolith suites which show clear evidence for derivation from the present lower crust of Archean cratons provide invaluable sight into the origin of the first emergent continents on Earth. Such xenoliths may bear fundamental information on the growth mechanism of the Archean crust, allow more accurate estimates of its total volume to be made, and may constrain the origin of the underlying mantle root and indicate whether it has remained coupled with the crust through the multi-phase geological history of the Archean craton.

2. Results

Twenty-two lower crustal xenoliths, mainly representing mafic garnet granulites, from the Kaavi–Kuopio kimberlites, have been studied in detail. Mineral thermometry *(Hölttä et al., 2000)*, together with isotopic and general petrological constraints imply that this sample suite is derived from the geophysically-determined, dense, high-velocity layer at the base of the crust (45–60 km deep). Single grain zircon U-Pb dates and Nd model ages (T_{DM}) both imply that this is hybrid layer consisting of both Archean and Proterozoic mafic granulites. Crystallisation ages of up to ~ 3.5 Ga equal to the oldest upper crustal ages *(Mutanen and Huhma, 2003)*, rendering it possible that thick continental crust existed in the Early Archean. Recrystallisation or dissolution-precipitation reactions of Archean zircons led to Proterozoic zircon dates that coincide with major events of basic magmatism within the craton. Most important post-Archean lower crustal growth took place due to ~ 1.9 Ga accretion of the Svecofennian arc complex to the craton margin when underplating basaltic magmas became mingled with Archean mafic lower crust. Later, at post-orogenic times ~ 1.80 Ga, lower crust became transiently heated, probably by magmas ponded at uppermost lithospheric mantle. Overall, Karelian lower crust records geological evolution of the craton through an epoch of over 3 billion years, implying that upper and lower crust have remained coupled since their origin in the Early Archean. Finally, our results are a new piece in the emerging data from several continents that emphasises the global importance of the 3.5 Ga episode. We suggest, that it represents a superplume event – more widespread than previously recognised *(Condie, 2001)* – when a significant fraction of Early Archean continental crust was formed.

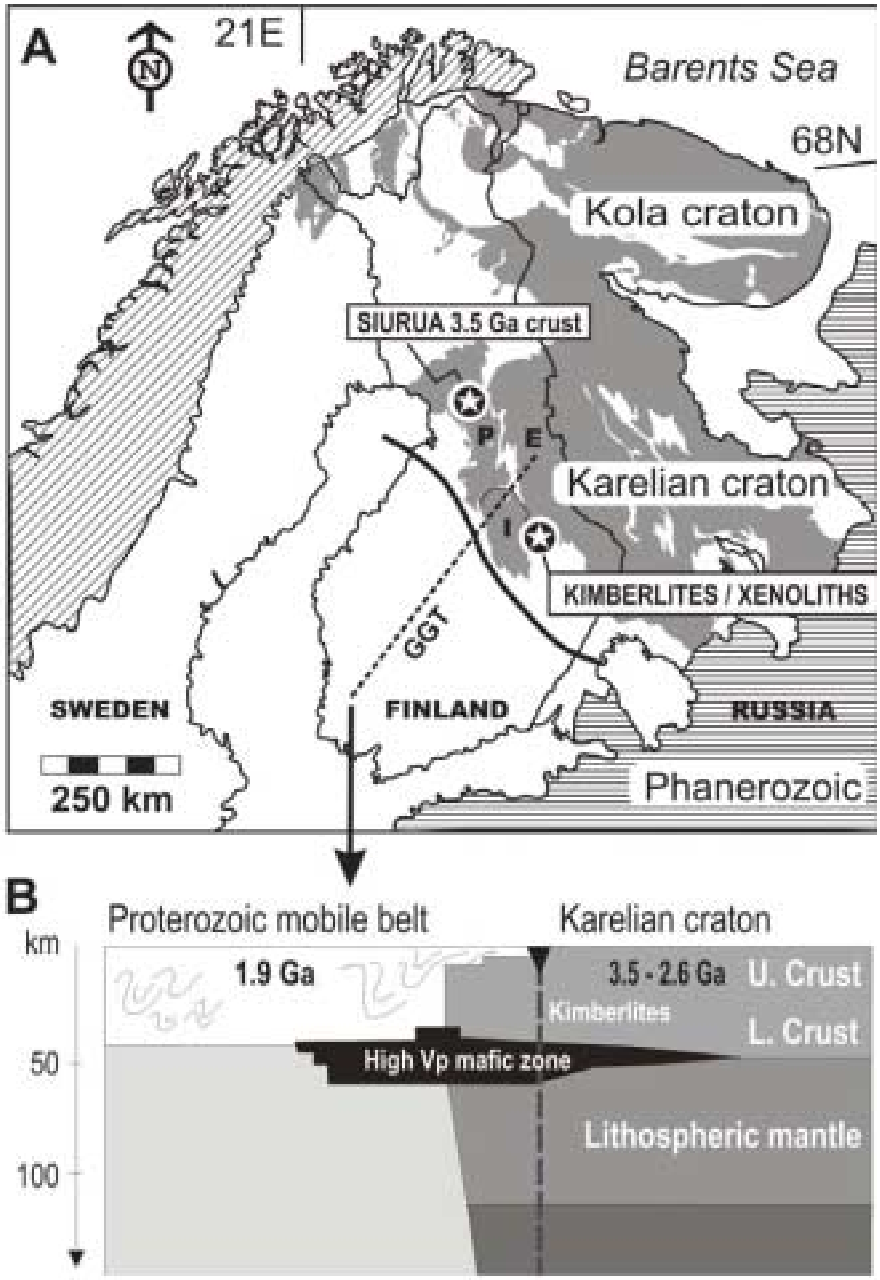

Figure 1. (A) Generalised geological map of the Fennoscandian Shield showing the location of the garnet granulite-bearing kimberlites, and the oldest known upper crust locality at Siurua. Shaded are is Archean granite-greenstone terrain and white is post-Archean, mainly Svecofennian rocks. Other abbreviations: Global Geoscience Transect -line (GGT); Iisalmi block (I), Pudasjärvi block (P), and Eastern Finland complex (E). (B) Cross-section along the GGT-profile (modified from *Korsman et al., 1999 and Lehtonen et al., 2004*). Note that only upper half of the lithospheric mantle is shown, its total thickness exceeding 240 km at this locality *(Kukkonen & Peltonen, 1999)*.

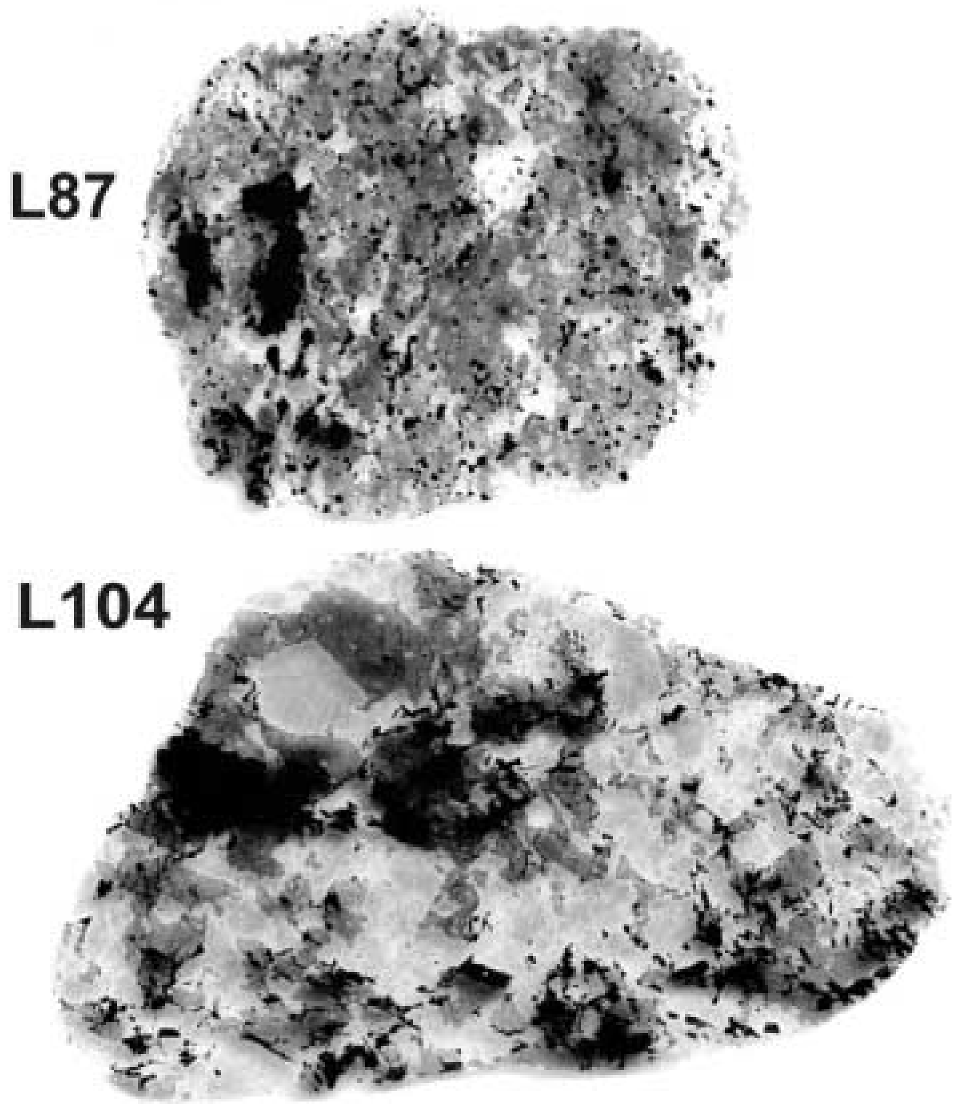

Figure 2. Examples of lower crustal xenoliths that have yielded Early Archean isotopic ages: (A) Mafic granulite xenolith L87, that has yielded U-Pb single grain zircon ages of > 3.5 Ga. (B) Two-pyroxene gneiss xenolith L104 that yielded neodymium model age $T_{(DM)}$ of ~ 3.7 Ga.

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3-D geophysical crustal model of Finland

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The 3DCM project aims to construct a 3D geophysical model of the crust beneath southern and central Finland using joint interpretation of seismic and potential field data. The project is carried out by the Department of Geosciences of University of Oulu and the Geological Survey of Finland (GSF). The main results of the 3DCM project are regional 3D P- and S-wave seismic velocity models and a regional 3D density model of the crust of southern and central Finland. In addition, new interactive software for large-scale (regional scale) 3D problems of joint interpretation of seismic and potential field data has been developed.

Keywords: Fennoscandian Shield, crust, lithosphere, joint inversion

1. On-going research: 3DCM project

The research project "*Three-dimensional crustal model of the Finnish part of the SVEKALAPKO research area*" (3DCM) is a joint venture between the scientists of the University of Oulu and the Geological Survey of Finland (GSF) who have worked in the SVEKALAPKO project and the Crustal Model Program.

The main objective of the 3DCM project is to construct a 3-D geophysical model of the crust beneath the SVEKALAPKO research area in the southern and central Finland. The model is based on the joint use of the seismic body wave data of local events registered during the SVEKALAPKO deep seismic tomography (SST) experiment, the potential field data (gravity and magnetic), digital geological maps, and the petrophysical database of the Geological Survey of Finland.

The model is required to improve inversion of SST and FIRE (Finnish Reflection Experiment) data and will advance the understanding of the tectono-geologic develop-ment of the Finnish bedrock.

2. The main results of the 3DCM project

At the moment, a regional 3-D density model of the SVEKALAPKO area has been created using joint interpretation of gravity data and existing seismic P-wave velocity model (Kozlovskaya et al, 2004a). The model comprises a volume of 1024 x830 x 72 km^3 and has been calculated on a 2 x 2 x 2 km^3 regular grid.

The 3-D P-wave and S-wave seismic velocity models of the crust of the SVEKALAPKO area were obtained from interpretation of the SVEKALAPKO local events data (Yliniemi et al., 2004, Kozlovskaya et al., 2004b). The models demonstrate the difference between the velocity structure of the Archaean and Proterozoic domains in the study area and good correlation of lateral velocity variations with major geological structures. In addition, the interpretation of local event data revealed several new details of the Moho boundary in the area and two groups of upper mantle reflectors beneath southern and central Finland at a depth of 70-90 km and 100-130 km (Yliniemi et al., 2004).

3. Software development

Compilation of the comprehensive large-scale 3-D geophysical crustal model demonstrated that existing techniques for 2-D and 3-D data processing (i.e. conventional filtration, gridding, interpolation etc.) are not sufficient in the case when the data are produced by various geophysical methods with different space resolution. The 3-D gravity modelling with the use of the existing petrophysical maps demonstrated that conventional interpolation of irregularly spaced data produces artefacts, which can result in misleading interpretations (Kozlovskaya et al. 2004a).

Therefore new gridding method and corresponding software (Petrock) has been developed to create improved maps of physical properties of bedrock of Finland from irregularly sampled data (bulk density and magnetic properties). The new interpolation method employed in the Petrock program uses the digital geological maps of Finland to provide additional weight to the interpolated and extrapolated values (Pirttijärvi, 2004a).

In addition, new interactive software for large-scale (regional scale) problems of joint interpretation of seismic and potential field data has been developed. The programs use parameterisation of 3-D distributions of geophysical parameters (seismic velocity, density, magnetisation etc.) by rectangular blocks (Pirttijärvi, 2003a,b,2004a,b,c). The Bloxer program (Fig. 1) allows interactive visualisation and handling of multi-parametric 3-D models. It contains many advanced features such as: editing of the parameter value and the size of the blocks, import and export of 3 D parametric data, dynamic memory allocation, and binary model files. The Grablox and Magblox programs are used for the forward modelling and inversion of gravity and magnetic field data with 3-D block models. Their inversion algorithm is based on the new linearised inversion method described in the PhD thesis of Pirttijärvi (2003c). The number of blocks used in the models is only restricted by the available computer memory. Moreover, in magnetic field modelling the inherent variation of Earth's external magnetic field can be taken into account. All programs are written in Fortran90 language for the MS Windows operating system. The visualisation and the user interface are based on DISLIN graphics library that allows the programs to be compiled and run also on Linux and Unix systems.

4. Further studies: enhanced 3-D geophysical crustal model of Finland

Further development of the crustal seismic model within the 3DCM project will be based on teleseismic receiver functions study (in collaboration with the Institute of Physics of the Earth of the Russian Academy of Sciences). The existing regional 3-D density model of the crust will be improved with the use of new map of the bedrock density of Finland and the data of FIRE seismic experiment (profiles FIRE1, FIRE2 and FIRE3). The model will be extended to the north using the new map of the bedrock density of Finland, the data of FIRE4 seismic reflection profile and results of previous seismic studies in northern Finland.

The mutual interaction of the 3DCM project with the Finnish Reflection Experiment (FIRE) experiment will improve the existing 3-D crustal model (in particularly, its uppermost part) and through the joint interpretation of the data of the FIRE project and potential fields will provide improved models in some key areas (for example, the areas of known ore deposits in Finland. Such interpretation will make it possible to investigate connection between reflectivity patterns produced by seismic reflection experiments and other physical properties (density and magnetisation), which is not completely understood in Finnish bedrock conditions.

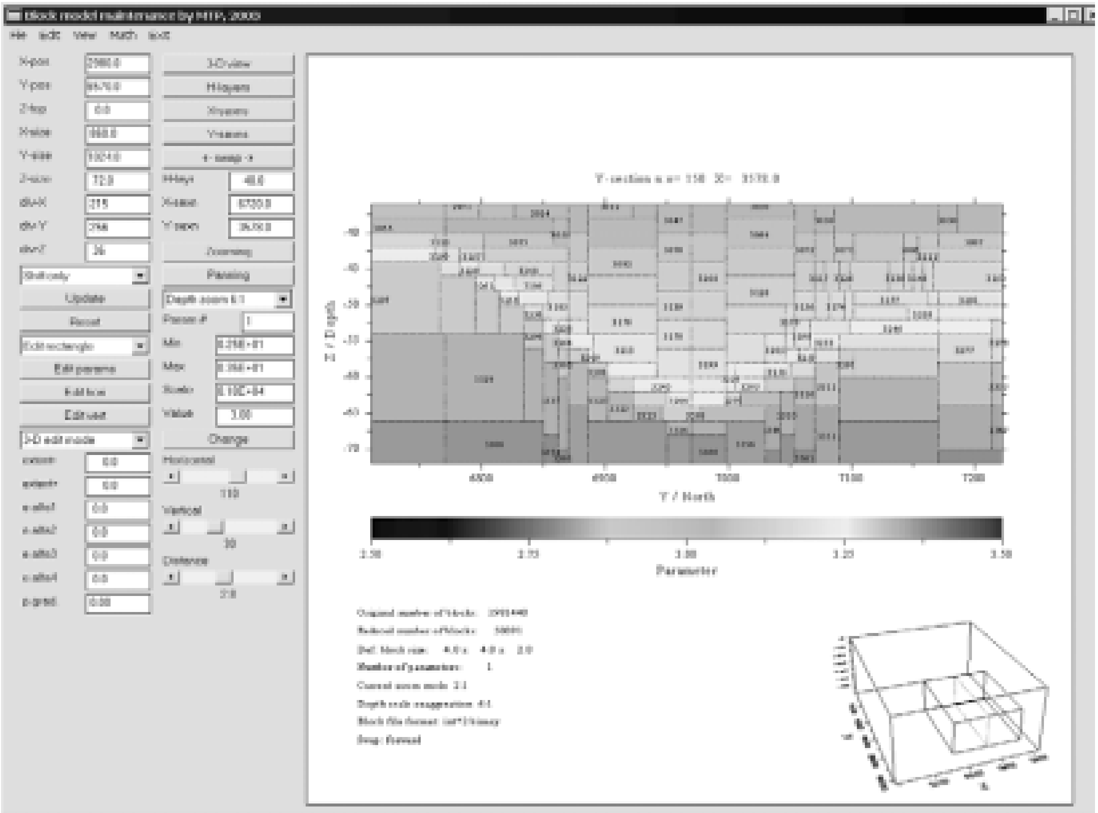

Figure 1. Screendump of the BLOXER program illustrating the graphical user interface and details of a vertical cross-section of the 3-D density model model (Kozlovskaya et al., 2004). The graphical user interface and the visualisation are based on DISLIN graphics library.

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The Lake Ladoga basin; preliminary insights into geochronology, igneous evolution, and tectonic significance

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Mesoproterozoic red-bed type sedimentary rocks, intercalated basalts, and subvolcanic mafic sills and dikes in the Lake Ladoga basin of Russian Karelia probably represent an intimate continuation of the magmatic events that led to the formation of the the *locus classicus* rapakivi granites in southeastern Finland and adjacent Russian Karelia in southeasternmost Fennoscandia. The mafic rocks of the basin are, in general, a CFB type, bear the Nd isotopic characteristics of ancient (Neoarchean) lithosphere, and yield a minimum age of 1460 Ma for the formation of the basin.

Keywords: basalt, dolerite, red-bed, basin, rapakivi granite, Mesoproterozoic

1. Geologic setting

The south-central part of the Fennoscandian (or Baltic) shield in Sweden, Finland, and adjacent Russia experienced widespread continental flood basalt (CFB)-type basaltic magmatism and red-bed sedimentation in the Mesoproterozoic (e.g., *Kohonen and Rämö, in press*). Published U-Pb zircon/baddeleyite data indicate ages of ca. 1260 Ma and ca. 950 Ma for this magmatism in southwestern Finland and central Sweden, respectively (*Suominen, 1991; Söderlund et al., 2004*). In the Lake Ladoga region, Russian Karelia, the Mesoproterozoic record registers the formation of a complex faulted-bounded basin, ca. 150 km in diameter, with intercalated sandstone, doleritic sills and dikes, and mafic lavas (e.g., *Amantov et al., 1996*). The mafic subvolcanic rocks are exposed on islands and skerries in the northeastern part of the basin (Valaam, Mantsinsaari, and Lunkulansaari islands) and basaltic lavas (eleven flows with an average thickness of about 15 m) are intercalated with clastic (mostly arkosic) sandstones in a section revealed by deep drilling in the Salmi-Tulemajoki region.

2. Geochemical observations

In general, the Lake Ladoga dolerites and lavas are alkaline, relatively evolved (< 6 wt. % MgO, 43–57 wt. % SiO_2), and strongly enriched in the light rare earth elements. The dolerites of the Valaam Island (and a gabbro from Lunkulansaari) have initial ε_{Nd} (at 1460 Ma) values between -9.6 and -8.6. This points to a Neoarchean, metasomatically enriched lithospheric mantle source (cf. *Upton et al., 1996*). Our preliminary Nd isotope data on the Tulemajoki basalts and basaltic cobbles on the coast of the Valaam Island have ε_{Nd} (at 1460 Ma) values of -5.8 to -4.1. These basalts may have been derived from a different source than the (more unradiogenic) subvolcanic mafic rocks.

3. Geochronology and tectonic implications

One concordant and one nearly concordant baddeleyite fraction from the dolerite on the Valaam Island in the northwestern part of Lake Ladoga yielded ^{207}Pb-^{206}Pb ages of 1459 ± 3 Ma and 1457 ± 2 Ma, respectively. This shows that the Mesoproterozoic CFB magmatism in the southeastern Fennoscandian shield commenced at least ca. 200 Ma earlier than previously anticipated and thus very shortly after the emplacement of the classic, 1670–1530 Ma (*Vaasjoki et al., 1991; Suominen, 1991; Amelin et al., 1997*), rapakivi granites. The basin formation probably started during the emplacement of the extension-related rapakivi intrusions as a result of thermal contraction of the rift zones, and the relatively thin crust hosting the rapakivi granites permitted ascent of the mantle-derived basaltic magmas. Whether the generation of these CFBs was related to a mantle plume is yet to be determined. In stratigraphy, the Mesoproterozoic basalts of the Lake Ladoga basin belong to the lower part of the basin (e.g., *Amantov et al., 1996*) and may actually represent basic volcanism related to the rapakivi event.

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Clay Mineral Occurrence in Prydz Bay Rise (ODP Site 1165), Antarctica: Implications for the Middle Miocene and the Middle Pliocene Ice Sheet Evolution

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The aims of this study are on the evaluation of climatic variability in the East Antarctic during 1) possible "warm" interval of the middle Pliocene and 2) the middle Miocene glacial event based on clay mineral occurrence. The results on the middle Pliocene with increasing smectite content and decreasing illite content may indicate dynamic conditions in the ice sheet behavior. The preliminary results on the middle Miocene sequence show possible indications of the climatic transition.

Keywords: Ocean Drilling Program, Antarctica, clay minerals, ice sheet evolution

1. Introduction

This study is a part of the larger Ocean Drilling Program (ODP) project, which represents the expertise in polar geosciences at the University of Oulu. From the beginning of 2004 the Integrated Ocean Drilling Program continued the work of ODP. It is an international organization that investigates global problems, especially in the field of geosciences. One of the major themes is the study of past climates.

The East Antarctic Ice sheet has existed for more than 33 Ma, but it has fluctuated considerably and has been one of the major driving forces of global sea level and climate throughout the Cenozoic era (Ehrmann and Mackenzen, 1992; Barret, 1999). Oxygen isotope records have been used to infer episodes of increased ice volume at 33.6 Ma, (Eocene/Oligocene boundary), 23.7 Ma (Oligocene/Miocene boundary), 12 to 16 Ma (middle Miocene), and 2.7 Ma (late Pliocene), the latter marking the onset of Northern Hemisphere glaciation (Flower and Kennett, 1994). However, data from other research proxies is also needed for better understanding of the Antarctic cryosphere and its relations to global climates and deep ocean circulation. It will also improve the evaluation of the long-term stability of Cenozoic ocean-climate system (*cf. Flower and Kennett, 1994*). The aims of this study are on the evaluation of climatic variability of the 1) "warm" interval of the middle Pliocene and 2) the middle Miocene glacial event based on clay mineral occurrence.

2. Regional setting and study material

Site 1165 was drilled on the Wild Drift (66°E and 79°E) on the Continental Rise off Prydz Bay, East Antarctica, to a total depth of 999.1 m below the seafloor (mbsf). The drilling recovered an extensive suite of terrigenous and hemipelagic sediments of the early Miocene to Pleistocene age *(Shipboard Scientific Party, 2001)*. One of the interests in this study is the high resolution sediment sequence between 0 and 50 mbsf which consist of a well-preserved section of Pliocene- to Pleistocene-age sediments. From this 50 meter sequence, a more intensely studied middle Pliocene from ~3.2 Ma (*cf. Whitehead and Bohaty, 2003*) was analyzed in 10 cm intervals. The other interest is from the middle to late Miocene sequence between 150 to 326 mbsf which is aged between ~10.0 to ~14.8 Ma *(Florindo et al., 2003)*.

3. Methods

X-ray diffraction (XRD) was performed on oriented clay samples as described by *Hardy and Tucker (1988)*. Diffractograms were recorded by a Siemens D 5000 with copper radiation (40kV, 40mA) at angles ranging from 2° to 32° 2ɗ (0.02° 2ɗ per second) immediately after the sample treatments. The principal four clay minerals smectite, illite kaolinite and chlorite were analysed by their basal spacings. MacDiff software version 4.2.5 (http://www.geologie.uni-frankfurt.de/Staff/Homepages/Petschick/RainerE.html) was used to quantify clay minerals. Since no internal standards were used, the precise accuracy of this procedure is not known, but the quantitative analyses justify interpretations of fluctuations around +/- 2%.

4. The middle Pliocene "warm" period

The East Antarctic Ice Sheet has a controversial glacial past. Some recent studies have shown that it may have varied from a polythermal, dynamic condition to a predominantly cold stable state as recently as the Pliocene time. Multiproxy study of the middle Pliocene could show the intervals of expansions of the ice sheet across the continental shelves and Antarctic cryospheric evolution, particularly during the middle Pliocene "warm" period 3.15 to 2.85 Ma *(cf. Dowsett et al. 1999; Murphy et al, 2002)* which may provide an indication of how Earth may respond to future global warming. The clay mineral results on the middle Pliocene with increasing smectite content and decreasing illite content may indicate warm and dynamic conditions in the ice sheet behavior.

5. The middle Miocene climatic transition

The middle Miocene represents the major phase of East Antarctic ice sheet expansion. The middle Miocene glacial event at ~14.8 – 14.2 is one of the largest climatic events during the Cenozoic. It occurred after the climatic optimum in the early middle Miocene (17 - 15 Ma) (Zachos et al., 2001) During the transition from ~14.8 to 12.9 Ma, climatic developments included major growth of the East Antarctic Ice Sheet and associated Antarctic cooling as well as a distinct increase in the meridional temperature gradient, large fluctuations in sea level followed by a global sea level fall, and important changes in deep water circulation, including increased production of Southern Component Water *(Flower and Kennett, 1994).* East Antarctic ice sheet growth and polar cooling also had large effects on global carbon cycling and on the terrestrial biosphere, including aridification of mid-latitude continental regions. *(Flower and Kennett, 1994).* The ODP Leg 188 Site 1165 provides record of the long-term lower to upper Miocene transition from a temperate climate to cold climate glaciation, with superimposed short-term glacier fluctuations since early Miocene time *(Shipboard Scientific Party, 2001).*The preliminary results on the middle Miocene sequence show possible indications of the climatic transition. Research of this kind, integrated with other palaeoenvironmental data may improve significantly our knowledge of the past Antarctic climate and our understanding of its evolution.

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Orogenic architecture and mineralization - general concepts, comparisons and relevance to Fennoscandia

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Interactions between orogenic architecture and thermal evolution are considered with respect to their role in mineral endowment. Genreic and specific examples of seismically defined architecture and mineralization are given, with emphasis on the Yilgarn craton and comparisons with Fennoscandia.

Keywords: seismic architecture, thermal evolution, gold mineralization, Yilgarn, Fennoscandia, numerical modeling.

1. Architecture of mineral systems

The importance of deformation and structural architecture in controlling fluid pathways and favourable depositional sites for hydrothermal ore deposits – particularly orogenic gold - is now widely recognized. Structural controls are evident across a range of scales and crustal depths, from pluton-centered hydrothermal systems in the upper crust, to plate-scale ductile shear zones systems channelling fluids during high-grade regional metamorphism.

In recent years, exploration strategies have been shifting away from empirical targeting of anomalies towards a more comprehensive understanding of the geological context of mineral systems, in an effort to define what makes a terrain prospective. Mass-balance constraints and metal enrichments in many hydrothermal mineral systems demand fluid access to large crustal volumes, and efficient transfer and focussing of fluids towards the ultimate site of deposition. Derivation of fluids from a deep crustal source implies the existence of some mechanism or process that enhances permeability in inherently low-permeability crystalline rocks. Moreover, the presence or absence of mineralization does not depend on favourable architecture alone. In addition, sustained fluid supply is required, as well as appropriate combinations of thermal, pressure and chemical gradients to optimize metal extraction, transport and deposition

In order to undertake such an integrated analysis of orogenic evolution, we need to

1. constrain both the 3-dimensional architecture of the terrain
2. document the thermal evolution, and nature and distribution of heat sources and transport, in order to establish the timing and nature of fluid and metal transport and deposition.
3. assess the role of the mantle in crustal orogenic evolution–

∉ is its effect dynamic in terms of its buoyancy and rigidity?
∉ does mantle heat flux control heat distribution and temperature-dependent rheology?
∉ does crustal magmatism and fluid flow involve material transfer from the mantle itself ?

We are then in a better position to consider the following issues:

∉ Can we identify critical structural architectures that are inherently favourable for the accumulation and preservation of mineral deposits?

∉ Is the thermal history and lower crustal composition conducive to efficient fluid tranpsort and focussing during deformation?
∉ Are there distinctive geochemical or isotopic signatures that link crustal granitic magmatism or sedimentary basin characteristics with a particular mineralization event?
∉ Does early cratonic history predetermine subsequent mineralization potential, in which case we have an effective tool in terrain selection, or can refertilization of mantle lithosphere lead to new endowment?
∉ Have secular trends in thermal evolution or geochemical differentiation (eg. composition and buoyancy of Archean versus younger lithospheric mantle) influenced orogenic processes?

Seismic surveys across a number of terrains on various continents and ranging in age from Archean to Cenozoic are now providing us with unprecendented opportunities for comparing orogenic evolution. Examples of well mineralised and prospective terrains for which seismic data are available include the Yilgarn Craton, Mount Isa Inlier, Curnamona Craton, Victoria, central New South Wales and Tasmania in Australia, and the Archean Abitibi belt (and other LITHOPROBE transects) in Canada, the Andes, Alaska, Sierra Nevada and the Witwatersrand Basin of the Kaapvaal Craton.

Examples of terrains that are not currently globally significant mineral producers, but whose seismic architecture and geodynamic evolution are relatively well constrained include the New England Fold Belt and Arunta Inlier in Australia, the Tibetan-Himalyan collision zone, the European Alps, New Zealand, Japan and the Archean and Proterozoic of the Fennoscandian Shield (Korja et al., 1993).

This presentation does not provide answers, but is intended to stimulate discussion on similarites and differences between Fennoscandian crustal architecture and a number of other terrains, with emphasis on implications for mineralization.

2. Different types of orogenic architecture

First order controls on orogenic archtiecture reflect a dynamic balance between externally imposed forces and the evolving internal thermal and material properties of the orogen. External boundary conditions include:
∉ rate of plate convergence
∉ duration of plate convergence
∉ direction of plate convergence
∉ mantle heat flux and thermal regime

Factors internal to the orogen include:
∉ inherited basement structure
∉ overlying basin architecture
∉ lithospheric composition, including compositional variatiosn controlling melting, and distribution of radiogenic heat production)
∉ lithospheric thickness, buoyancy and elevation
∉ lithospheric anisotropy

Coupling between crustal and lithosphere during convergence has essentially two principal geometries, as illustrated in Figure 1.

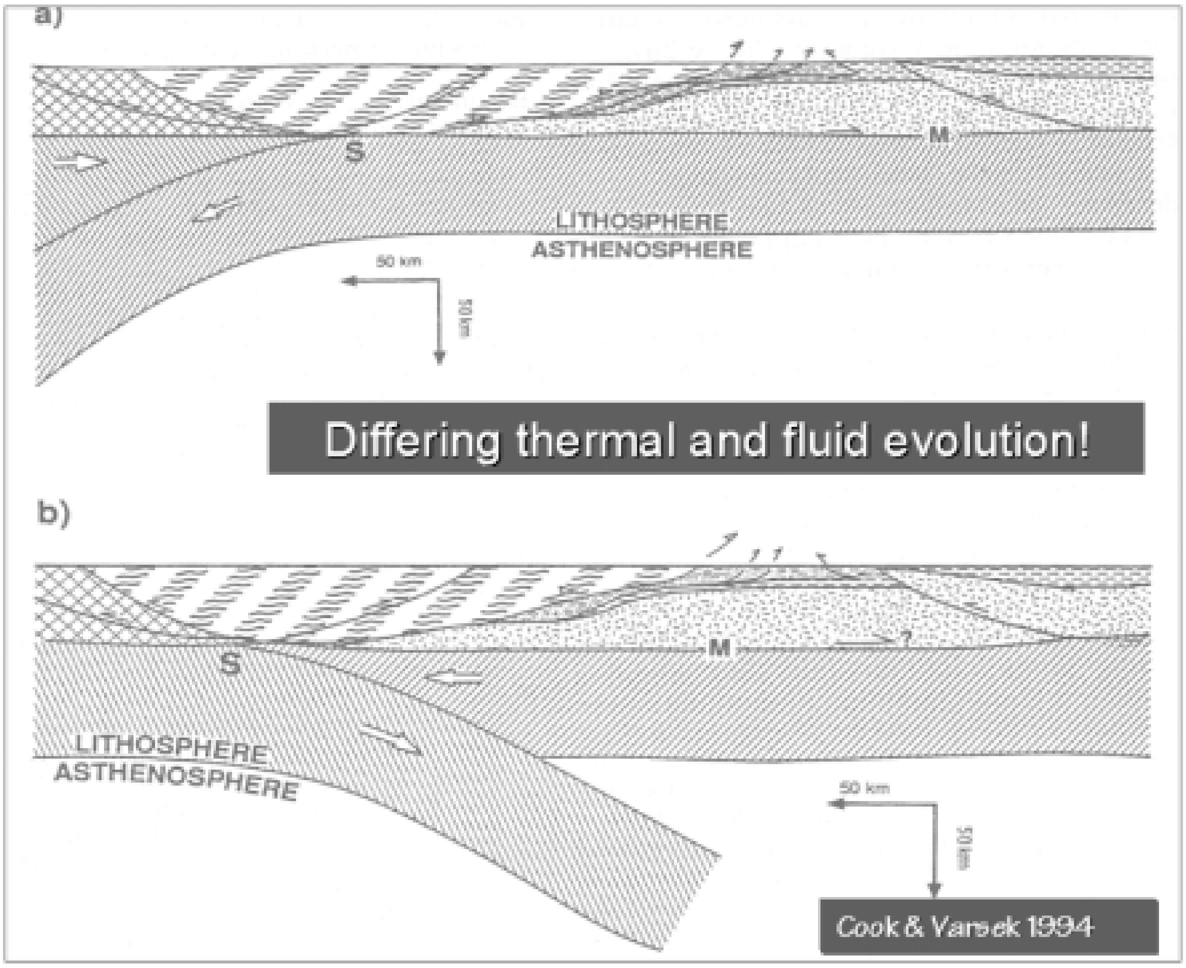

Figure 1. The two fundamental collisional orogenic architectures – subduction away from the orogen, and subdcution beneath the orogen (*Cook and Varsek, 1994*).

Two general types of fold belt profiles may be distinguished – (1) asymmetric, foreland migrating wedges with various combinations of imbricates, duplexes and antiformal stacks, and (2) doubly vergent profiles, with either tectonic wedges, backthrusts and triangle zones, or conversely, divergent thrusts leading to pop-up wedges. A critical feature in all of these architectures is, by analogy with traps in petroleum systems, the potential exists for creating fault-bounded domains of differential uplift and overpressuring beneath relatively impermeable units that act as seals to contain hydrothermal systems.

Because many gold deposits occur in vein systems, where quartz-dominated lithologies define the brittle-ductile transition is also interesting to speculate whether there is a thermal and rheological feedback that maintains fluid supply and focussing in critical state, a characteristic of self-organizing systems (Bak, 1996).

3. Asymmetric, foreland migrating fold and thrust belts

A foreland-propagating fold and thrust belt will have a different thermal evolution and uplift history depending on rate and duration of convergence. This architecture is evidently favourable for fluid migration and formation of relatively low temperature zinc mineralization in deforming sedimentary basins but may be less favourable for orogenic gold however, as the typical clockwise P-T-t evolution means that thermal equilibration in the overthrust terrain is attained after peak deformation. However, this settings is typical for porphyry Cu-Au deposits formed from pluton-driven hydrothermal systems in uplifting terrain in mamgmatic arcs. However, because such deposits are associated with uplift in elevated terrain, their long term preservation potential is likely to be low, except by a combination of rapid subsidence and orogenic collapse and temporary burial beneath a foreland basin.

4. Doubly vergent deformation and tectonic wedging

Doubly vergent orogeny (*Koons 1990*) such as described from the Southern Alps of New Zealand have the form of large divergent, pop-up wedges, with the potential to provide uplift in combination with efficient fluid transfer, including deep penetration of meteoric fluids (*Craw et al. 2001*). This is one mechanism for addressing the problematic issue of early metamorphic dehydration and the relatively late timing of many orogenic gold systems.

It is also common for divergent architectures to develop by late stage backthrusting. The thermal and tectonic evolution associated with such retrowedge formation has also been analyzed numerically (*Jamieson et al. 1998*); it appears that after initiation of a retrowedge, tectonic imbrication in the middle and lower crust can develop with opposite vergence to that in the overriding wedge. These features are variously referred to as tectonic wedges or triangle zones, which are commonly developed during folding and thrusting in inverted sedimentary basins (Jamison 1993). This architecture is of particular interest in that it has the potential to generate effective permeable seals during orogenic deformation; this may then enhance focussing, circulation and retention of fluids within a middle crustal environment, compatible with greenschist facies conditions during mineralization. The process may also lead to an effective decoupling of lower crustal and upper crustal deformation styles late orogenic isostatic readjustment may provide a greater opportunity for exhumation and preservation of greenschist facies crustal levels.

This type of architecture and history can be inferred from seismic and geologic data from the late Archean Yilgarn craton (*Drummond et al. 2000; Sorjonen-Ward et al., 2002*) as shown in Figure 2. It is also likely to be relevant to the evolution of Lapland and the context of gold mineralization, where the bivergent nature of deformation has long been recognized . Although the polarity of orogeny in the Yilgarn is not so well defined as in Lapland, with no obvious foreland, the architecture and thermal regime at the time of mineralization are better constrained.

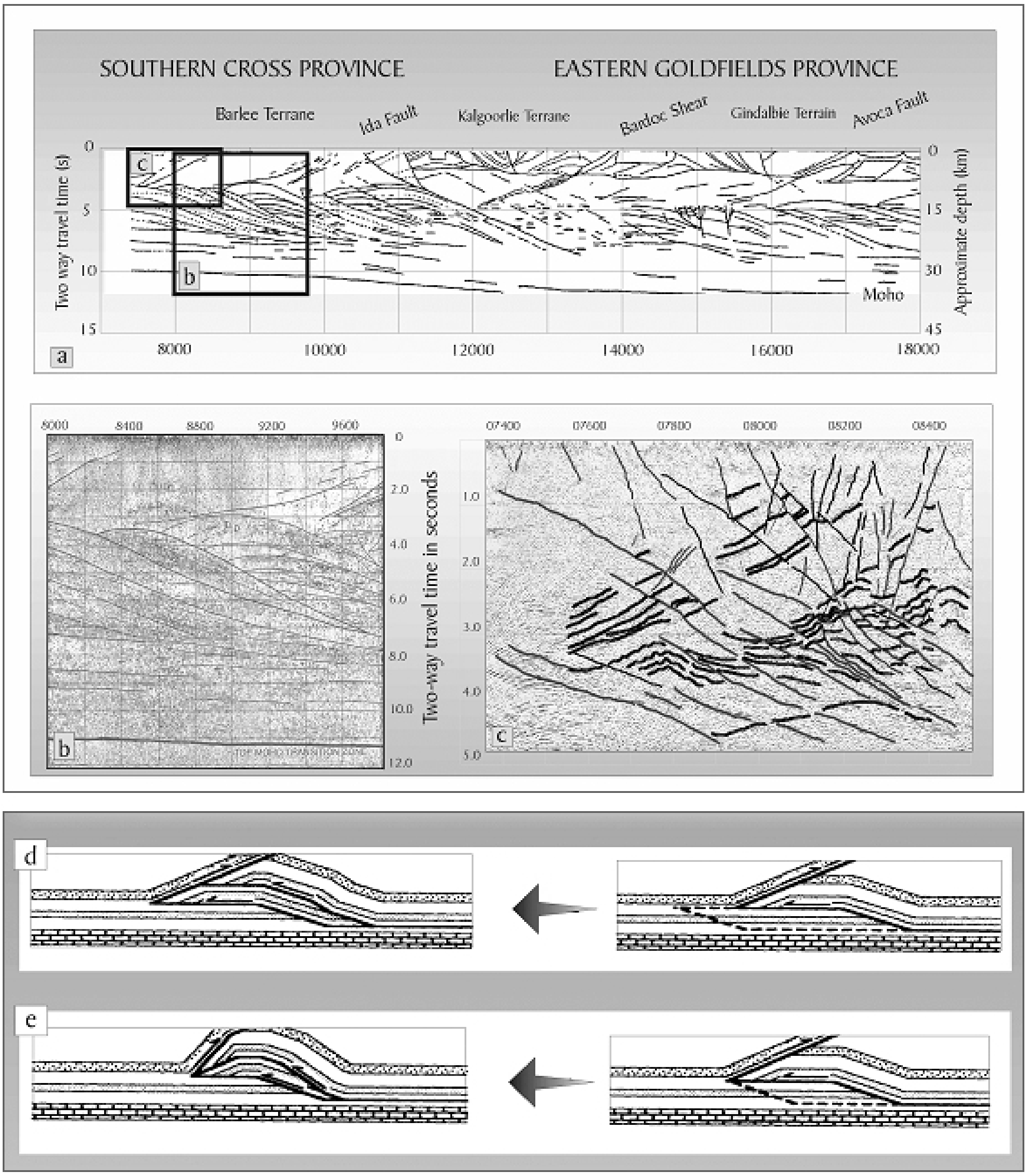

Figure 2. a-c: summary sketches of principal seismic reflectors in Eastern Goldfields Traverse EGF-01, from Drummond et al. (2000), and Sorjonen-Ward et al. (2002); d, e: general concept of tectonic wedging (after Jamison 1993).

It appears from the regional context of mineralization that most Yilgarn gold deposits formed at a relatively late stage in the structural and thermal evolution of the craton, perhaps 30 Ma or more after the earliest deformation events. Structural studies of vein systems favour a compressive rather than extensional deformation regime during mineralization. It is therefore reasonable to infer that mineralization occurred during mineralization. It is therefore concluded that mineralization occurred during active uplift and exhumation but during ongoing compression. The seismically defined detachment zone may record an effective mechanical decoupling between the upper crust, which deforms in a very heterogeneous manner, accompanied by uplift and erosion, and the middle and lower crust, which displays a distinct asymmetric tectonic imbrication. Indeed the sesimic data strongly suggest that middle crustal deformation zones facilitated melt transfer from the lower crust to higher levels, giving rise to sheeted granitic complexes at the base of the greenstone sequences, at conditions approximating the brittle-ductile transition. Sequestering of fluids by lower crustal melting and release during crystallization is one mechanism for supplying fluids to the upper crust late during orogenesis, consistent with the observed temporal overlap between mineralization and granite emplacement, and potentially buffering the redox character of magmas and fluids. Results of numerical modeling of fluid flow during deformation (*Sorjonen-Ward et al. 2002*) also illustrate the importance of lateral fluid flow driven by asymmetric thrust migration and loading and topographic elevation and the ability of "pop-up" wedges to facilitate fluid upflow and downflow during uplift.

The backthrusting or tectonic wedging geometry identified in the Yilgarn is also discernible in a number of other deep seismic sections through orogenic belts of varying age, including highly mineralized terrains such as the Paleozoic Lachlan fold belt in Western Victoria, (which contains the giant Bendigo and Ballarat gold deposits) and the Mesozoic Great Valley sequence of the Sierra Nevada (hosting the Mother Lode). A somewhat different architecture, where a post-collisional arc has developed over a rifted foreland can be identified in the Cambrian Mount Read volcanics of Tasmania In many respects this is similar to the architecture and evolution of the Skellefte district of northern Sweden; by inference, this type of scenario needs to be considered in interpreted the seismic data in western Finland, as well as the Tampere district.

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IODP investigates the solid earth's cycles and global environmental change

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Finland has a membership in the Integrated Ocean Drilling Program (IODP) through the European Consortium for Ocean Research Drilling (ECORD). IODP uses multiple drilling platforms to accomplish the goals of the Initial Science Plan. It has three major themes: 1) The deep biosphere and the subseafloor ocean; 2) Environmental change, processes, and effects; and 3) Solid earth cycles and geodynamics.

Keywords**:** Earth's dynamics, climatic change, resources and deep biosphere

1. IODP 2003-2013

IODP builds upon the legacies of the Deep Sea Drilling Project (1968-1983) and the Ocean Drilling Program (1983-2003). IODP is an international scientific research initiative that began its scientific work in 2004. European Consortium for Ocean Research Drilling (ECORD) is formed through a joint action of the European scientific community together with funding agencies presently from fourteen nations to provide a single European-based entity in IODP. The Academy of Finland supports the Finnish membership in IODP. ECORD's role is to co-ordinate Europe's participation in IODP. ECORD in structured into 4 components parts: ECORD Council, Managing Agency (EMA), Science Support and Advisory Committee (ESSAC) and Science Operator (ESO).

Overall IODP program management is provided by IODP Management International, Inc. (IODP-MI) having offices in Washington, D.C. and Sapporo, Japan, and is responsible for program-wide science planning, oversight of engineering development, publications, education and outreach, site survey data management, and core sample depositories. IODP-MI receives advice from the international IODP Science Advisory Structure (SAS).

IODP explores the history and structure of the earth as recorded in seafloor sediments and rocks. IODP will use the drill ship JOIDES Resolution for its first expeditions. In addition, IODP will utilize a riser vessel currently under construction in Japan named Chikyu ("Earth") to achieve program's science goals. European countries (ECORD) will contribute mission specific platforms (MSP) for drilling in shallow water (less than 30 m) and ice-covered regions. MSPs are necessary to investigates earth's regions and processes that where previously inaccessible or poorly understood such as Arctic. The first MSP operation was taken place August 7 to September 16, 2004. The Arctic Coring Expedition (ACEX) investigated Arctic climate history by drilling along the Lomonosov Ridge. The cores contain evidence of a dramatic defrostering of the Arctic Ocean near the North Pole 55 million years ago and a long, slow slide towards the perennial ice cover of recent times (*Kerr, 2004*).

2. IODP's scientific themes

IODP is science-driven. The initial science plan describes the goals of the program and has three major themes: 1) The deep biosphere and the subseafloor ocean; 2) Environmental change, processes, and effects; and 3) Solid earth cycles and geodynamics.

Eight initiatives are identified that are ready to be addressed within the next decade of drilling: 1) Deep biosphere; 2) Gas hydrate; 3) Extreme climates; 4) Rapid climate changes;

5) Continental breakup & sedimentary basin formation; 6) Large igneous provinces; 7) 21st century moholes; and 8) Seismogenic zone.

3. IODPFinland.oulu.fi

This portal is planned to support Finnish membership in IODP. It will provide basic information and important links to those web sites that relate to the program. This site is especially planned for easy start for Finnish scientists and students to look for opportunities for participation for coming expeditions, find educational resources or to begin planning or jointing to international drilling proposal initiatives. Participation is open to scientists and engineers (professors, research scientists, technologists, graduate students) from all IODP member countries.

Responsibility of updating the national IODP/ECORD portal for information and announcements is for Finnish ESSAC delegates. To support Finnish membership in IODP also a group of experts from universities and surveys is established.

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Integration of geophysical data with sediment composition, examples from Prydz Bay (ODP Site 1166), Antarctica

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Integration of geophysical and sedimentological data can improve the reliability and accuracy of data interpretation. For example P-wave velocity and density values are very informative when used alongside with geological data.

Keywords: Ocean Drilling Program, Prydz Bay, Antarctica, geophysical and sedimentological data

1. Introduction

The Ocean Drilling Program (ODP) was a broad international scientific program, which worked in the field of Earth Sciences and it has continued IODP phase from the beginning of 2004 (*Strand et al., 2002*). ODP and IODP data are used for studying the structure and evolution of the Earth. In addition to this, programs provide a good possibility to study past climatic conditions and variations. Finland participates in IODP as a member of the European Consortium.

The ODP depositories and databases include samples from seafloor sediments and rocks as well as geophysical *in situ* –borehole measurements and sediments' physical properties. These databases are Janus database (available in the web: http://iodp.tamu.edu/janusweb/general/dbtable.cgi) and LDEO's database (http://www.ldeo.columbia.edu/BRG/ODP/DATABASE/index.html).

In this work there has been used material from ODP Leg 188, Site 1166, which is located on the continental shelf in Prydz Bay, East Antarctica. The used data consist petrophysical and geophysical borehole logging measurements, concrete samples and thin sections. This paper is based on Tiensuu's M.Sc. thesis (*2004*).

Different data types were studied together and it was considered, whether certain geophysical data have characteristics, which could be useful in determining the boundaries and in identifying sediment or rock material in the specific units. Correlating geophysical measurements with mineralogical and sedimentological data has been the basic idea in this work, thus aid to improve also sedimentological interpretation. (*Tiensuu, 2004*)

2. Some notifications of data integration

Geophysical measurements can improve the quality and reliability of the available data. Usage of a wide range of different geophysical logging methods together with petrophysical and geological methods, and coupling their results together may solve the problems caused by poor core recovery, which is often the case in ocean research.

3. P-wave notices

P-wave velocity seems to be one of the most promising petrophysical parameters in order to solve the state of compaction and sedimentological history. One benefit of seismic velocities is that they can be measured with logging methods and when the core recovery is sufficient, also petrophysically.

4. Density notices

The use of petrophysical density data is more restricted by core recovery problems than seismic velocity data and it also does not tend to be sensitive enough for small variations. Not even logging data is very sensitive for small-scale variations. Density data measured with logging methods are more continuous and usually of better quality than petrophysical data, due to the fractioning of actual samples during drilling.

High density values suggest relating with high clay fraction content. Naturally the used methods also affect the results. Bulk density mean and median values measured with logging tool HLDS were regularly higher than those of petrophysical results, and also the standard deviation was in a remarkably lower level. Higher densities are due to that in logging techniques measured material does not break, meanwhile in conventional petrophysical research some cracks are formed during the drilling. These cracks form more easily into the core than to the surrounding.

5. General notices

Generally, it seems that standard deviation values remain remarkably lower in logging data than they do in petrophysical data. We see that as an indication of better data quality, something expectable due to higher sampling frequency and less destructive (less crack-forming) methods. In case of high-quality geological samples also petrophysical methods can provide significant data. The best combination would be utilizing geophysical logging methods alongside with different geological, mineralogical and geochemical methods.

This study has verified the assumption that geophysical measurements can increase and augment the available data. Geophysical data can be used at least for making educated guesses for material type, even if there are no concrete samples, which is often the case in borehole studies. Usage of a wide range of different geological, petrophysical and geophysical logging methods and coupling their results together can bring some ease to the problem caused by poor core recovery.

6. Conclusions

In conclusion, the best combination would be utilizing geophysical logging methods alongside with geological (petrographical and geochemical) methods. In case of high-quality geological samples also petrophysical methods can expand the general view of research subject and provide significant new details; thus improving the validity of the conclusions. Regarding to the glacial history of Prydz Bay, the integration of geophysical data with sediment composition, has proven to be a useful approach. However, apparent correlations may not always imply a causal relationship, and relations of different variables are often complex. Therefore, caution is recommended when drawing final conclusions from the integrated data. Many times, however, more reliable interpretations can be made by integrating sediment compositions with measured geophysical properties at the same intervals. Especially geophysical logging data can be used to infer sediment and rock properties in intervals with low core recovery.

There is also a need for further studies of the subject together with statistical analysis. More detailed mineralogical and petrological studies combined with petrophysical and *in situ* –measurements could give valuable information on mineral content's relation to physical properties. Especially the mineral content, studied by e.g. electron microprobe and reflected light microscope (ore microscope), of the opaque minerals could be interesting.

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The Anjalankoski earthquake swarm in May 2003

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Composite focal mechanism of the May 2003 earthquake swarm in Anjalankoski, SE Finland corresponds to dip-slip motion along a nearly vertical fault plane. The epicenters of the 16 earthquakes can be associated with the internal intrusion boundary of the Vyborg rapakivi batholith. The events are most likely occurring at depths between 1 to 2 km below the surface. The unusually shallow swarm activity is characteristic to the Vyborg rapakivi area in southeastern Finland.

Keywords: Earthquake swarm, relative location, focal mechanism, Finland

1. Introduction

During May 2003, the Finnish seismograph network recorded a swarm of 16 earthquakes in Anjalankoski, SE Finland (Fig.1). In spite of the small size of the events (ML=0.6-2.1), the swarm was widely felt in the area surrounding Anjalankoski. Rattling of windows and small dishes accompanied by explosion- or thunder-like sounds were typical descriptions of the felt effects. These effects and excitation of a strong Rg phase on seismograms recorded at epicenter-station distances up to 150 km suggested that the focuses were most likely very shallow.

Historical reports revealed that most of the macroseismic epicenters in figure 1 also represent double or multiple earthquakes, accompanied by similar felt effects as the Anjalankoski swarm. The duration of individual swarms varied from several days to several years, the time gap between the largest subsequent swarms being ca. 50 years. The best known historical swarm occurred in Lapinjärvi, about 40 km SW of Anjalankoski. A total of 23 earthquakes were reported felt there during August 1951 and October 1956. The Anjalankoski swarm is, however, the first one verified by instrument recordings.

2. Geological and geophysical background

The bedrock of southern Finland has been formed in the Svecofennian orogeny between 1.9 and 1.8 Ga. The regional scale lithological units and prominent shear zones are striking SW-NE. The Svecofennian lithologies have been intruded by rapakivi granite between 1.65 and 1.50 Ga. The earthquake swarms occur within the Vyborg rapakivi batholith consisting of number of small intrusions. The intrusions take an elliptical form with an elongated axis in 060^{o} direction (Fig.1; *Härme, 1980*). This is also the main strike direction of the regional magnetic and Bouguer anomalies. Main directions of vertical joints are 060^{o} and 330^{o} (*Simonen, 1987*). Vertical fault surfaces with fluorite and calcite fillings also occur in 060^{o} direction in the Loviisa area (*Sundblad and Alm, 2000*)

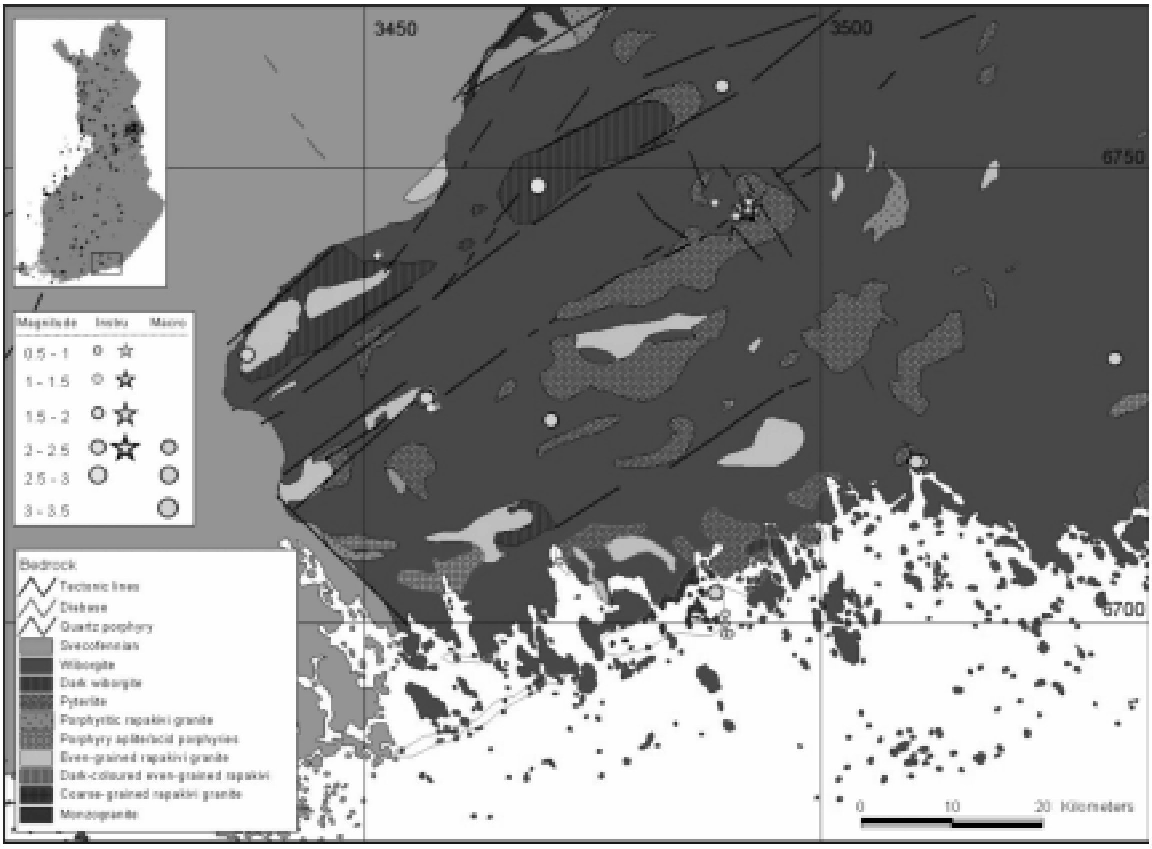

Figure 1. The epicenters of southeastern Finland on a lithological map modified after *Härme (1980)*. The Anjalankoski earthquakes are marked by asterisks. Inset map in the upper corner shows regional seismicity.

3. Data and event processing

Waveforms from eight short-period and three broad-band stations of the Finnish seismic network were examined to determine source parameters of the earthquakes. The epicenter-to-station distances varied between 40 km and 360 km. The distribution of stations was not optimal (azimuth gap 110^{o}-$250\ ^{o}$), which affected both the hypocenter location and the source mechanism inversion.

However, the waveforms within the swarm were nearly identical and thus it was possible to apply relative earthquake location method to better define the geometry of the cluster and to identify the fault associated with the activity. Relative hypocenter location method HypoDD by *Waldhauser and Ellsworth (2000)* was used to relocate the cluster. At the nearest three stations, time-domain cross-correlation method was applied to pick automatically the P and S arrival times with an accuracy needed for relative location. The rest of the arrival times were picked manually. Optimal velocity model for the Anjalankoski area was constructed on the basis of refraction and earthquake data.

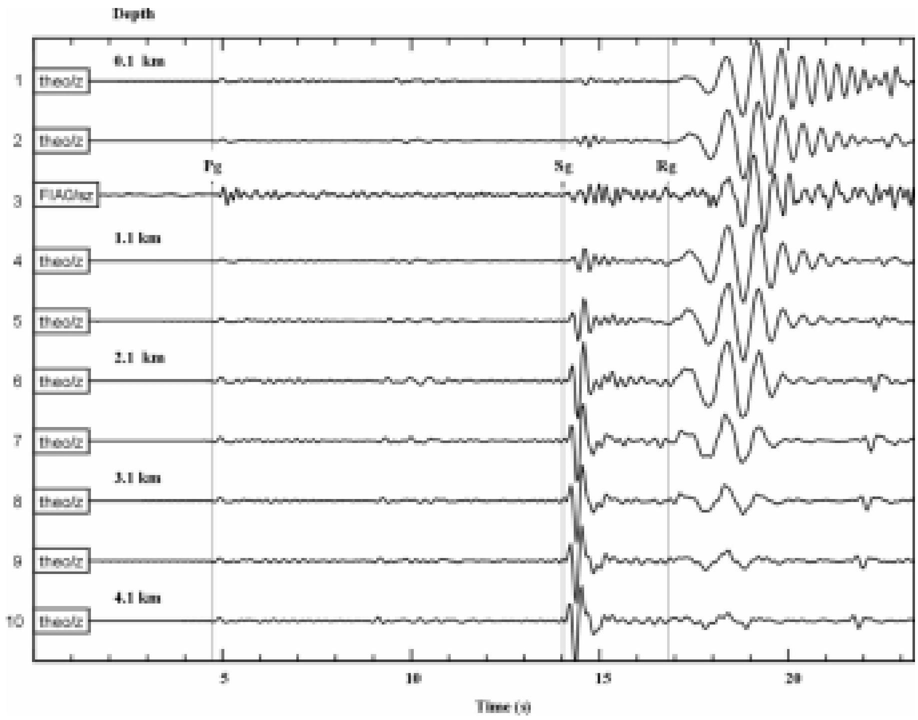

Figure 2. Vertical component record of an Anjalankoski earthquake at station FIA0 (D=80 km) at Sysmä. Synthetic section shows the excitation of Rg phase relative to body wave phases (P, S) as a function of source depth. Synthetic seismograms have been calculated using the composite focal mechanism and the local velocity model. Note that Rg is not well developed for sources deeper than 2 km.

Composite fault plane solution of the five strongest earthquakes was determined using P-wave polarity data together with SV/P and SH/P phase amplitude ratios. Synthetic waveform modeling (Fig. 2) was used to further constrain and verify the source and structural parameters. The methods were described in detail in *Uski et al. (2003)*.

4. Results and conclusions

- Both the historical and the instrumental earthquake data indicate that unusually shallow swarm activity is characteristical to the Vyborg rapakivi area in southeastern Finland.
- The composite fault plane solution corresponds to dip-slip motion along a nearly vertical or a horizontal fault plane.
- The nodal plane striking $252^{o}/82^{o}$ coincides with an internal intrusion boundary of the Vyborg rapakivi batholith.
- The relocated epicenters are also aligned along a SWW-NEE trending zone.
- Relative location as well as synthetic waveform modeling of the Rg wave suggested that the events have been unusually shallow, most likely occurring 1 to 2 km below the surface.

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Carbonate vein-dykes in Naantali: a new occurrence of carbonatite in southwestern Finland?

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A group of carbonate vein-dykes has been discovered in the city of Naantali, southwestern Finland. Textural, mineralogical and chemical evidence is consistent with a magmatic origin. It is here proposed that the vein-dykes in Naantali constitute a new discovery of carbonatite surrounded by an area of potassic fenitization.

Keywords: Carbonatite, Naantali, Finland

1. Introduction

A group of at least 15 carbonate vein-dykes has been identified in the city of Naantali in southwest Finland. The vein-dykes range in width from 1-60 centimetres, and intrude Svecofennian enderbite. In a wide area around the vein-dykes, the host rocks have subsequently been altered, characterised chemically by removal of silicon and addition of potassium. The altered area also contains widespread randomly oriented calcite - epidote veins as well as pervasive staining from iron oxides. Presented here is chemical and textural evidence that the carbonate vein-dykes have a magmatic origin.

2. General Geology

Vein-dykes are found in an area extending for approximately one kilometre in length and 300 metres in width. Outcrops are, however, mostly restricted the steep western slopes of Kuparivuori or to roadcuts. Extreme sensitivity to weathering, soil or water cover, and the presence of man-made structures inhibits the search for additional vein-dykes. The majority of the vein-dykes follow the same general NW-SE trend and dip to the northeast at an average angle of 45°. The vein-dykes cut the existing schistosity of the surrounding rocks, the difference in strike being 60-90° and in dip 45-50°.

Carbonate rocks are light grey to pink in colour, and often show a banded or layered texture. The vein-dykes are composed of medium to fine grained calcite, with repetitions of grain size variation parallel to the margins. Iron oxides are common as an interstitial dusting in streaks also parallel to the strike. Similar texture from the Wasaki complex of Kenya is taken as evidence by Le Bas (1977) of flow parallel to the margins and crystallisation from molten magma. In addition to the iron oxides, accessory minerals visible in hand specimen include apatite, quartz, allanite, and chlorite. Andersen (1984) describes a similar paragenesis from the Rødberg portion of the Fen complex. In one outcrop, fragments of the wall rock appear within the vein-dyke and appear to have been plucked from the sides. Given the low viscosity of carbonate melts (e.g. Dobson et al. 1996), this would require a forceful intrusion and rapid cooling.

The vein-dykes intrude dark grey Palaeoproterozoic (1,88 Ga; Väisänen & Hölttä 1999) Svecofennian enderbite. A petrography of these rocks is given by Helenius (2003). Alteration surrounding the vein-dykes can be divided into three zones. All three zones contain abundant, randomly oriented calcite − epidote veinlets, whose width rarely exceeds 1mm. Zone I is closest to the vein-dykes, and is equivalent to a dark red syenite or monzanite. Zone I is typically less than 1 metre wide around each vein-dyke. Zone II is approximately 500-600 metres in width and is pink to red quartz syenite or quartz monzanite. Zone III extends an additional 200-300 metres from the vein-dykes, and is a pyroxene bearing grey granodiorite, with narrow bands (≤1cm) of Zone II type alterations around calcite + epidote veinlets. The original schistosity of the rocks is preserved in Zone III, but is no longer evident in Zone II. Grain size increases with intensity of alteration from 0,5-1,5 mm in Zone III to 3-5 mm in Zone I.

3. Petrography

Thin section examination reveals that the texture of the carbonate rock is hypidiomorphic. Calcite is subhedral and comprises approximately 90% of the rock. Quartz occurs as fine-grained aggregates and appears limited to areas near the contact. Fine grains of apatite, allanite, chlorite, fluorite, and titanite occur as accessory minerals. Opaque minerals occur generally in bands, and some mineral grains, particularly apatite, are stained with iron oxides. SEM examination has revealed monazite and members of the bastnäsite series as inclusions in apatite.

Allanite is non-metamict, strongly pleochroic, and commonly forms haloes around aggregates of apatite and quartz. Woolley et al. (1991) describe this texture from extrusive carbonatites in the Uyaynah area of the UAE. Allanite is not a common mineral in carbonatites, but it has been reported from various locations including Bayan Obo, China (Yang & Le Bas 2004), Mountain Pass. California and Uyaynah, UAE (Woolley et al. 1991).

Zone I altered rocks one I contain K-feldspar, plagioclase, chlorite, and blue-green alkali amphibole. Silica in the form of free quartz has been almost completely removed from the rock. Feldspars are heavily altered and pervasively stained with iron oxides. Chessboard albite, as described by Kresten & Morogan (1986) from the Fen complex also occurs. Calcite, apatite, clinopyroxene, epidote and opaque minerals occur as accessory phases.

Zone II rocks are similar. however, plagioclase is more abundant than K-feldspar. Zone II rocks also contain more quartz and less alkali amphibole than Zone I. Alteration is less pronounced in feldspars, but iron oxide staining is still present throughout. Epidote, clinopyroxene and opaque minerals also occur as accessories, but apatite is no longer an important phase, and calcite only occurs in the calcite + epidote veins.

In Zone III, alteration of the same type as in Zone II occurs only in bands (≤1cm) around epidote + calcite veins. Biotite can be seen reacting to chlorite in the outer areas of these bands. Outside of these bands, the rock contains plagioclase, quartz, K-feldspar, biotite. and orthopyroxene (enstatite). It should be noted that although the grey rocks of this zone appear unaltered, K-feldspar is present, while it is not present in the unaltered rocks examined by Helenius (2003). Microscopic-scale quartz veinlets also occur in this zone.

4. Geochemistry

Whole-rock geochemical analysis was conducted using X-ray Fluorescence (XRF) at the Geological Survey of Finland. Table 1 shows the analytical results as average concentrations of selected elements for the vein-dykes and each of the alteration zones. Enderbite analyses for background are taken from Helenius (2003).

Table 1. Average compositions (XRF) of rocks in the study area. Enderbite compositions taken from Helenius (2003).

	Vein-Dyke	Zone I	Zone II	Zone III	Enderbite
Na2O %	0.02	1.71	5.74	5.27	3.94
MgO %	0.86	4.34	2.5	2.25	3.15
Al2O3 %	1.3	14	15.4	15.58	15.8
SiO2 %	9.68	61.25	67.55	67.31	63.36
P2O5 %	1.13	0.27	0.21	0.19	0.22
K2O %	0.18	5.65	2.06	1.96	1.58
CaO %	49.06	5.32	1.59	2.98	4.58
TiO2 %	0.02	0.58	0.49	0.47	0.62
MnO %	0.09	0.09	0.06	0.05	0.09
Fe2O3 %	0.78	3.99	4.01	3.49	6.3
Sc	55	-	-	-	-
V	-	94	75	66	72
Cr	-	77	78	66	81
Ni	-	44	51	38	50
Rb	-	140	85	62	74
Sr	2016	1636	850	914	587
Y	44	14	-	-	-
Zr	-	122	135	137	138
Ba	43	1819	340	349	292
La	644	91	-	-	-
Ce	1411	215	45	56	60
Th	12	-	-	-	-

Carbonate rocks show enrichment in Sr, La, Ce and Y. Zone I alteration is characterised by enrichment in Mg, Ca, K, Sr. Ba. La and Ce. Zone I rocks are depleted in Si and Na. Zones II and III show elevated levels of Si, Na, K, Sr, and Ba.

Three samples were analysed for stable carbon and oxygen isotopes. Sample JW0404, from a vein-dyke, was analysed at the University of Helsinki; samples PSH-04-01.1 from a vein-dyke and PSH-04-01.2 from a calcite + epidote veinlet in Zone II were analysed at the Geological Survey of Finland. Results are given in Table 2.

Table 2. Carbon and Oxygen isotope ratios. See text for sample descriptions.

	JW0404	**PSH-04-01.1**	**PSH-04-01.2**
δ13C PDB	-11.48	-11.29	-5.99
δ18O SMOW	9.96	11.7	-16.89

5. Discussion

Carbonatites are igneous rocks comprised of greater than 50% carbonate minerals (e.g. Le Bas 1977). Many different criteria have been proposed for the identification of carbonatites, unfortunately no single piece of evidence is truly diagnostic. Evidence from multiple sources including textures, mineral assemblages, bulk chemistry, alteration, and stable isotopes must be considered.

Carbonatites are characterised by high levels of Sr, Ba, LREE and Y (e.g. Puustinen & Karhu 1999). The vein-dykes in Naantali are enriched in Sr, Y, Ce and La. Ba, though virtually absent in the vein-dykes, is highly enriched in altered rocks, particularly in Zone I. Although carbonatite melts are typically high in K and Na, low-alkali carbonatites are known, for example at Fort Portal, Uganda (Barker & Nixon 1989). The Naantali vein-dykes lack alkalis, however Zone I is enriched in K and Zones II and III are enriched in both K and Na.

Fenitization is a metasomatic process occurring around carbonatite and alkaline igneous intrusions typically involving the removal of silica and the addition of alkali elements (e.g. Le Bas 1977). Fenites show a great deal of variation in chemical composition, based on variations in the compositions of the fenitizing fluid, degree of fenitization, and original composition of the rock. Vartiainen and Woolley (1976) distinguish between sodic and potassic fenitization, while Kresten (1988) makes divisions of contact, aureole, and vein-type fenites. In Naantali, addition of K and removal of Si corresponds to potassic fenitization. Zone I alteration in Naantali represents contact fenitization; Zones II and III are aureole fenites of decreasing intensity.

Special consideration should be given to the relative mobility of certain elements. Rollinson (1993) describes elements including K, Sr, Rb, and Ba as mobile, while P, REE, Sc, V, Y, Ti and Zr are immobile. In the Naantali vein-dykes, mobile elements characteristic of carbonatites are enriched, but enrichment is diluted over the entire area of alteration. Immobile elements present in the vein-dykes, such as P, Y, REE, and Sc are not found far from the vein-dykes, while immobile elements present in the enderbite, such as V, Ti, and Zr are found at constant levels throughout the altered rocks. This, combined with the widespread epidote – calcite veinlets and staining from iron oxides, is interpreted as evidence of late-stage alteration by an oxidizing fluid. Whether the fluid source was degassing of the carbonatite or a later event is not known.

A $\delta^{13}C$-$\delta^{18}O$ isotope comparison is often used to trace the origins of limestones. $\delta^{18}O$ values from the vein-dyke samples fall within the normal range for carbonatites, while the value from the veinlet is only slightly heavier. Combined with the $\delta^{13}C$ values, the vein-dyke samples fall just below (lighter C) the field for carbonatites given by Deines & Gold (1973) whereas values from the veinlets fall to the right (heavier O). Isotope values are within the range of -2 to -12 $\delta^{13}C$ PDB and +10 to ~26 $\delta^{18}O$ SMOW given by Barker & Nixon (1989) for the Fort Portal carbonatite in Uganda. Values are also close to those given by Puustinen & Karhu (1999) for the Halpanen carbonatite in southeastern Finland.

Carbonate vein-dykes in Naantali show magmatic textures including flow banding and incorporation of wall-rock fragments on an outcrop scale. Chemical evidence is consistent with established normal values for carbonatite. Isotopic evidence is also consistent with other carbonatites. This combined body of evidence is here interpreted to mean that the carbonate vein-dykes in Naantali do constitute a new discovery of carbonatite.

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